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A TEXTBOOK OF GEOLOGY

TEXTBOOK OF GEOLOGY

PUBLISHED BY

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Physical Geology. By the late L. V. Pirsson. Third Edition revised by Wiliam M. Agar, Assistant Professor of Geology; Alan M. Bateman, Professor of Economic Geology; Carl O. Dunbar, Associate Professor of Historical Geology; Richard F. Flint, Assistant Professor of Geology; Adolph Knopf, Professor of Physical Geology; Chester R. Longwell, Professor of Geology. Revision Edited by Chester R. Longwell. Cloth; 6 by 9; 488 pages; 322 figures.

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Frontispiece

Eruption of Halemaumau, Kilauea Volcano, Hawaii, May 24, 1924. This "cauliflower cloud" was considerably more than a mile high when the photograph was taken. (Maehara, Hilo, Hawaii.)

A TEXTBOOK OF
GEOLOGY

PART I — PHYSICAL GEOLOGY

BY

LOUIS V. PIRSSON

Late Professor of Physical Geology in the Sheffield Scientific School of Yale University

PART II — HISTORICAL GEOLOGY

BY

CHARLES SCHUCHERT

Professor Emeritus of Paleontology in Yale University and of Historical Geology in the Sheffield Scientific School of Yale University

PART I

THIRD EDITION, REVISED BY

William M. Agar, Assistant Professor of Geology
Alan M. Bateman, Professor of Economic Geology
Carl O. Dunbar, Associate Professor of Historical Geology
Richard F. Flint, Assistant Professor of Geology
Adolph Knopf, Professor of Physical Geology
Chester R. Longwell, Professor of Geology

Revision Edited by Chester R. Longwell

Second Printing Corrected

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PREFACE TO PART I, THIRD EDITION

The textbook of geology by Pirsson and Schuchert appeared fourteen years ago, and the first revision of Part I was completed shortly before Professor Pirsson's death in 1919. Both parts of the text have been indorsed by numerous universities and colleges; but teachers who know Pirsson's volume from repeated use have come to feel the need of important changes in it to bring the subject matter up to date and to strengthen the treatment of various topics. At the request of Professor Schuchert, author of Part II of the textbook, six members of the Department of Geology in Yale University have coöperated in the present thorough revision. The revisers have attempted to preserve the excellent balance for which the earlier editions have been commended. There has been no desire to make the book more elementary, but the aim has been to clarify the treatment so far as possible. With this end in view the revisers have not hesitated to recast statements, to transfer, add, or omit large sections, or even to rewrite entire chapters. The total length is essentially unchanged, although the number of pages is slightly increased because a uniform size of type has been adopted.

The older editions of the book used a two-fold division into "Dynamical Geology" and "Structural Geology." From their experience the revisers do not favor this division as a teaching device, as it necessitates a somewhat rigid order of subjects and leads to some awkwardness in the treatment. For example, earthquakes can be explained to best advantage after the student has considered the effects of crustal movements as seen in structural features; and some aspects of landscape development cannot be appreciated fully without knowing the nature and the scale of folds, faults, igneous bodies, and numerous other things that ordinarily are classed as structural. A proper understanding of geology comes only with the realization that the Earth is dynamic and changing. It appears that this viewpoint is presented to best advantage by considering the results of past activity along with the analysis of dynamic processes. This method gives a unified picture which is partially lost by any deliberate separation of the *dynamic* from the *static* aspects of the Earth.

No claim is made that the order of presentation chosen for this edition is the best possible. Probably there will always be considerable differ-

ence of opinion in this matter, because the subjects treated in geology are so interrelated that some anticipation and repetition is unavoidable with any order of arrangement. Of necessity the study begins with some consideration of minerals and rocks; but as laboratory materials are required, and as most instructors use some kind of laboratory manual dealing with these materials, no introductory chapter on minerals and rocks is included. Some of the common minerals are described briefly in an appendix, and as methods of studying the minerals are also explained this appendix may serve as a short manual for the student or for the interested layman. Descriptions of the common igneous, sedimentary, and metamorphic rocks appear in the chapters that give systematic discussions of these rock groups.

The following conspicuous changes appear in the book as revised: The introductory chapter is completely rewritten. Discussion of the larger features and relations of the Earth is transferred from former Chapter X to the new Chapter II. The chapter on the atmosphere is rearranged, with the idea of emphasizing soil formation instead of wind erosion; and a short discussion of weather and climate is added. The treatment of stream erosion is considerably amplified; the erosion cycle is emphasized, and new sections explain erosion in semiarid and arid climates. Glaciers and glaciation are discussed directly after streams, and ground water is placed before lakes and swamps. The former chapter dealing with the geologic rôle of life is omitted and its diverse subject matter appears in the treatment of weathering, swamps, and the ocean. At the suggestion of Professor Schuchert the chapter on the ocean has been expanded by the transfer of considerable material from Volume II of the text. The discussion of sedimentary rocks is presented as soon as all the agencies responsible for their formation have been considered. Igneous rocks are discussed directly after volcanoes. Folding, faulting, and warping are brought together in one chapter, and the discussion of earthquakes follows. The treatment of metamorphism has been changed to make it accord with present conceptions. A new chapter on land forms deals with the more complicated aspects of river history, explains the sculpturing of mountains in various stages of their development, and seeks to give the student some appreciation of landscape as related to geology. The discussion of ore deposits has been simplified considerably. A geologic time table is appended because the revisers have found it advisable to give students a general acquaintance with the time scale before they begin the formal study of historical geology. An annotated list of references is added to each chapter for the use of those who may want further reading on any subject. The geologic map of North America issued with the older editions is omitted.

The part performed by each of the revisers is indicated below.

Chapters 1, 12, 13, 15, 16, by Longwell
Chapters 2, 10, 11, 14, by Knopf
Chapters 3, 5, 18, by Bateman
Chapters 4, 7, 17, by Flint
Chapter 6 and Appendix A, by Agar (now of Columbia University)
Chapters 8 and 9 and Appendix B, by Dunbar.

As the revision progressed the work of each man had the benefit of criticism by his colleagues; and finally the entire revision passed through the hands of a single editor.

An attempt has been made to improve the illustrations as well as the text. Block diagrams are used freely to explain the development of surface forms and to illustrate structural features. Philip B. King executed most of these diagrams. Many halftones have been replaced by new views, and the total number has been increased somewhat. Individual credit is given for photographs used; but the revisers wish to mention their special obligation to the U. S. Geological Survey, the U. S. Army Air Corps, and the Hawaiian Volcano Observatory.

Chester R. Longwell.

New Haven, Conn.
January 15, 1929.

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PHYSICAL GEOLOGY

CHAPTER I

THE SCOPE AND METHOD OF GEOLOGY

Man has a natural curiosity about the Earth, his home. Of what materials is it made? When and how did it come into being, and through what changes has it passed? What has been the story of life on the Earth, and exactly what part has man himself played in the drama? Inquiries such as these have engaged the attention of thinking people from a very early period, as evidenced by the mythologies of the ancients. Some of the old philosophers — Pythagoras, Aristotle, and others — caught brilliant glimpses of the truth; but the science of geology, which strives for the full answer to questions about the Earth and the record of life, had its first consistent development in modern times, with the growth of science in general.

Any understanding of the Earth must begin with some knowledge of the substances that compose it. We are permitted to see only a thin rind of the globe, and therefore all information about the vast interior portion must depend on indirect evidence. However, our most active interest lies in the part that can be explored directly; and this part in itself provides an almost limitless field of study. It is a zone composed of rocks and their constituent minerals. Even to the casual observer it is evident that these materials, exposed in cliffs or in road cuts, tunnels, and other artificial excavations, are highly varied in character; and accordingly it may appear that only a specialist can hope to gain any adequate acquaintance with rocks. Fortunately it has been found that comparatively few types are important quantitatively in the visible part of the Earth, and therefore any educated person may learn without great difficulty to recognize most of the rock masses he may see in the Alps, the Rocky Mountains, or elsewhere in his travels. A study of variations in rock types or of the more detailed features in minerals must of course be left to men specially equipped for the task.

The practical value of recognizing one rock from another is readily apparent from the viewpoint of the professional geologist or the mining engineer. In the search for petroleum and for ore minerals, in the se-

lection of sites for great dams to make storage reservoirs, or in constructing an intricate subway system for a large city, knowledge of rocks and their peculiarities is a vital necessity. But there is a much broader interest in the subject, and the general key to this interest is within easy reach. A visitor to the slopes of Vesuvius or of Mauna Loa sees masses of dark, slaggy rock. It is fairly obvious, even without actually seeing fluid lava, that this dark material was once liquid and flowed down the slopes as a red-hot stream. Continued investigation would convince the traveler that the fire-made or *igneous* rocks are common in many lands and have been formed at many different dates; they constitute an important part of the bedrock beneath us. Again, a brief examination in parts of the high Andes or Himalayas reveals layers of compacted sand or mud which include an abundance of sea-shells. Phenomena of this kind recall the conclusion of Aristotle, " The relation of land to sea changes, and a place does not always remain land or sea throughout all time."

Thus the grouping of rock masses according to the mode and place of their origin is not merely a dull classification for the convenience of scientists; it is the first step in the fascinating game of unraveling the ancient history of a region. In all of its many aspects, the study or practise of geology recognizes this fundamental interest in past events. But just as the proper understanding of human history requires some knowledge of present day social, economic, and political conditions, so the deciphering of events in the geologic past is dependent on acquaintance with processes still operating on and within the Earth. Rocks are not inert monuments to conditions and forces that no longer exist. At active volcanoes we may observe all stages in the development of new rocks from molten material (Fig. 1). In the muds of river deltas and the oozes on sea floors we find the modern equivalents of ancient beds, now greatly distorted and eroded, which furnish the shells and other fossils so common in high mountains and plateaus. Rivers, glaciers, and other surface agencies are etching and slowly wearing away the lands. In some continents the land is being lifted up at a measurable rate, and there is evidence that some mountains are in process of active growth. Thus we are actual witnesses to a constant struggle between Titans; some that work at the surface, striving to tear down the rocky continents, and others within the Earth that persistently oppose the leveling process. Rocks are being destroyed and others are forming to replace them. Activities which can be seen and analyzed have been persistent throughout geologic time, and accordingly we may use the present as a key to the past. All processes now engaged in modifying the Earth are important in the preliminary study of geology.

From the geologist's viewpoint, therefore, the Earth is dynamic and changing; not static and inert. However, since most of the changes are very slow as judged by human standards, it is necessary to form some conception of geologic time in order to appreciate the continuity of Earth-history. The study of geology revolutionizes ordinary notions of time, just as a moderate knowledge of astronomy gives a new vision of space. Our solar system has grand dimensions, and yet in its entirety it is but a point by comparison with the stupendous diameter of the

Fig. 1. — Igneous rock in the making. The dark-colored rock is solidified lava; the white band is a stream of fluid lava, flowing toward the observer. (The Alika flow, 1919. Hawaiian Volcano Observatory.)

starry galaxy. Similarly we think of the earliest human records as very ancient; but in a geological sense the first appearance of man is a modern event. The oldest relics of primitive men are found only in the superficial soil or other rock débris formed in the latest geologic epochs. In all the time that has elapsed since the oldest civilizations existed, the major landscape features of the Earth have remained essentially unchanged; but we know from the immensely longer geologic record that generations of mountains were made and worn away, before any evidence of man appeared, by the same deliberate forces now at work. The length of time required for such transformations has been almost inconceivably great. How long has it taken for the Colorado River to excavate the Grand Canyon? We are appalled at the realization

that the slow action of water has carried away so many cubic miles of solid rock. Nevertheless the Grand Canyon is a youthful feature. The general record of earlier events is written plainly in the rocks traversed by the trail in climbing the vertical mile from the inner gorge to the outer rim (Fig. 2). Mountains were made and worn down; then the land was submerged beneath the sea for long ages, and continued to sink slowly while limy ooze, mud, and sand built up a deposit thousands

Fig. 2. — A minor canyon tributary to the Grand Canyon. Colorado River shows at lower right. The horizontal layers of rock represent accumulations on sea floors during long geologic periods. The cutting of the canyon by stream erosion is a recent event geologically, although it required a long time measured by human standards. (U. S. Army Air Corps.)

of feet thick; this loose material was converted into firm rock; and finally the rising of the wide plateau region high above the sea permitted the cutting of the canyon. Minimum estimates fix the age of the lowest rocks in the gorge at hundreds of millions of years; and yet these rocks may be youthful in comparison with the total age of the Earth. The closest study of the geologic record has revealed "no traces of a beginning, no prospect of an end."

This broad, philosophical aspect of geology rests on a secure foundation only because of patient effort by generations of workers in all lands.

Prior to the nineteenth century many natural philosophers formed their ideas on geological subjects largely by deductive reasoning, in ignorance of the facts to be learned by observation in the field. Gradually it came to be realized that inductive study, based on all the facts obtainable, is essential for any safe conclusions. A vast array of field evidence has been accumulated. The great mountain ranges have been fruitful fields for geologic investigation, since they furnish the finest exposures of rock formations. But no sources of information are neglected. Geologic features everywhere are examined closely in the field and represented accurately on maps. Mines give opportunity to explore beneath the surface, and deep wells drilled for water or oil yield valuable data. Modern instruments devised to record earthquake waves and to measure the value of gravity have made it possible to draw important conclusions regarding the invisible interior. Slowly but surely the Earth is giving up many of its secrets to the inquisitiveness of man.

Of necessity a field so broad as the study of the whole Earth calls for a division of labor among specialists. Some workers give their principal attention to the rock formations laid down in former seas, lakes, and rivers; some to the volcanic and other igneous rocks; some to the mineral veins and other deposits of economic value. Other subjects of special study are the deformation of rocks by folding and fracturing; the various land forms sculptured by surface agencies; the fossils entombed in rocks; and the minerals that make up rocks of all kinds. All groups of investigators recognize unsolved problems, and by their united efforts continue to widen the frontiers of the science as a whole. Like any other growing subject, geology extends beyond the lighted zone of proved fact. It has also a large twilight zone of inference and probability into which the full light of investigation continues to spread slowly; and beyond this a region of shadow and complete darkness, relieved only by scattered flashes of speculation. The speculative side of the subject is in some respects the most fascinating, but it may also be dangerous to the uninitiated. A comprehensive discussion of geology takes account of fact, inference, and speculation, but distinguishes carefully between them. Careless broadcasting of hypotheses as if they were proved facts has given rise to numerous popular misconceptions about the Earth and its life.

The study of geology begins logically with an introduction to the important kinds of rocks and minerals. It proceeds to an examination of the forces that act on the outer part of the Earth, and the changes produced by these dynamic agencies. Finally, the keys provided by this preliminary study are used to explore the long geologic record contained in the rocks.

CHAPTER II

A GENERAL VIEW OF THE EARTH

The Earth and Its Neighbors. — The Earth is one of a group of planets that revolve around a common central orb — the sun. Some of these, like Jupiter, are much larger than the Earth; some like the asteroids, or minor planets, are much smaller; some are much nearer the sun, others farther away. The group has nearly a common plane of revolution about the sun, as suggested in Fig. 3, and this fact is thought to have an important bearing on the origin of the solar system. The Earth and other planets were born of the sun; according to the modern theory of cosmogony, which we owe to the American geologist, Chamberlin, the substance of the planets was expelled from the sun under the influence of the disruptive pull of a passing star several times larger than itself; but how this ejected matter was aggregated to form the planets is a problem on which ideas are still far apart. The common plane of revolution of the sun's satellites is possibly an inheritance from the biparental origin of the system. The birth time of the Earth was probably 2000 million years ago, and since that time the Earth has been gradually acquiring its present character. A vast span of time has thus elapsed, and that portion the record of which is written in the rocks is referred to as " geologic time."

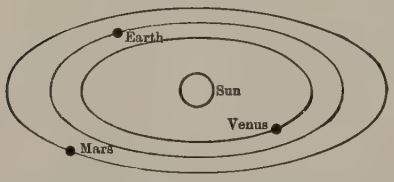

Fig. 3 —Planets revolve about the sun in nearly one plane, as suggested by the orbits of three of them.

The sun is called by astronomers a G-type dwarf star, by which they mean that it has already lived three-fourths of its life as a self-luminous body. Its surface temperature is approximately 6000° C. as determined observationally; the temperature at its center is estimated to be 40,000,000°. The geologic record shows that the sun has been supplying light and heat to the Earth at an approximately uniform rate during hundreds of millions of years. Up to the early part of the present century the other sciences could not adequately explain how the sun had been able to maintain this prodigal expenditure of energy. It is now considered probable that its fires are stoked either by utilizing sub-

atomic energy or by the conversion of matter into radiant energy — the annihilation of matter. According to Eddington, the transmutation of the elements would keep the sun going 10,000 million years, but the annihilation of matter would give " a very ample time scale."

The mass of the sun is 332,000 times that of the Earth. Even counting in the masses of the other planets, practically all of the mass of the solar system is in the sun. Therefore the sun is the overwhelmingly dominant member of the system. Carrying with it the Earth and the other planets, the sun is travelling through space at the rate of 12 miles a second.

The path of the Earth about the sun is not a circle, but an ellipse, one of whose foci is the sun. The deviation of the ellipse from a circle, however, is relatively small; the average distance of the Earth from the sun is nearly 93 million miles. In consequence of its elliptic orbit, the Earth is 3 million miles nearer the sun on January 1 than it is on July 1. This fact causes the summers to be somewhat cooler and the winters somewhat warmer in the northern hemisphere than they would otherwise be. At the present time summer, as measured from the vernal equinox on March 21 to the autumnal equinox on September 23, is 7 days longer than winter. In the southern hemisphere, however, winter is the longer season. Owing to the precession of the equinoxes, this condition will be reversed between the two hemispheres 10,500 years hence. Because of the perturbing effect of the other planets, the eccentricity of the Earth's orbit varies during a period of 500,000 years, and at maximum eccentricity the Earth is 13 million miles nearer the sun in summer than in winter, thereby causing short hot summers and long cold winters. At that time the winters will be 35 days longer than summer. As we shall see later, these astronomic conditions have been held by some to be sufficient to produce great climatic changes, to determine cycles of sedimentation, and even to bring about the several glacial epochs that have affected our planet during its long career.

Besides revolving around the sun, the Earth is spinning on its polar axis, each rotation in 24 hours giving rise to day and night. The axis of rotation is not perpendicular to the plane of the Earth's orbit but inclined to it at an angle of $66\frac{1}{2}°$, and this inclination gives rise to the seasons, summer and winter, alternately in the northern and southern hemispheres according as the axis is pointed toward the sun or away from it.

The Earth, as we have seen, is a very insignificant fraction of the solar system, and for that matter the solar system itself is but an insignificant fraction of the universe. The sun is a modest star in a stellar system whose members are numbered in thousands of millions. At immense distances beyond this system — our " universe, " as it is cur-

rently called in astronomy — are other gigantic systems. According to Sir J. H. Jeans, "the farthest astronomical objects whose distances are known are so remote that their light takes over 100 million years to reach us." Nevertheless, throughout this vast extent, with its island-universes containing myriads of stars in various stages of development, the same general physical laws that we know on the Earth appear to govern. Gravity operates in the same manner; light is transmitted everywhere by the same kind of vibrations; the spectroscope tells us that the same chemical elements occur in distant stars as on our Earth. Moreover, the meteorites that our planet gathers in its journey through space and which appear to be the disrupted fragments of former worlds are made of substances identical with those found on the Earth. Consequently there appears to be a unity of law and a uniformity of material throughout space, and we feel justified in assuming that facts and reasoning derived by astronomical study of the other heavenly bodies may also be applied in our study of the Earth.

Form of the Earth. — The Earth is not a true sphere but a spheroid flattened at the poles, so that the axis on which it rotates is slightly shorter than the equatorial diameter. The polar flattening (oblateness) of the Earth and the other planets is fully explained by the principles of celestial mechanics. It depends on the centrifugal force due to rotation and on the internal constitution of the planet. By internal constitution is meant the composition, density, and distribution of the density within the planet.

It is now accepted that the rate of rotation is gradually decreasing, or in other words that the day is growing longer, at the rate of $\frac{1}{1000}$ second per century. The reason advanced for this diminution is that the fluid friction that is caused by tidal turbulence in the shallow seas on the borders of the continents uses up energy, and as this energy is derived chiefly from the Earth's energy of rotation it therefore retards the rate of rotation. Two-thirds of the frictional loss occurs in Bering Sea, a shallow body of water having strong tidal currents. As the size, depth, and number of such seas has varied greatly throughout geologic time, in fact a large part of Historical Geology is the record of the expansion and contraction of such seas, it is plain that the rate of retardation cannot have been constant.

On the other hand, it is probable, though not proved, that the Earth has been contracting through loss of heat; because of the resultant shrinkage in size it would rotate faster. This increase might well counterbalance the retarding effect of the tidal friction. The problem of the net increase or decrease of the Earth's rotation during geologic time must therefore be set down as unsolved.

RELIEF FORM OF THE EARTH

General Features. — The irregularities of the Earth's surface, or its relief, divide naturally into major and minor groups. The major relief features are the continents and the ocean basins. The continents are regarded as being partly submerged beneath the oceans, and the submerged portions are known as the continental shelves. The minor relief features of the continents are the mountains as opposed to valleys and basins; and the minor relief features of the ocean basins are islands and submarine ridges as contrasted with the profound troughs or *deeps* in the ocean floor.

The average height of all the lands above sea-level is 2300 feet. North America averages about 2000 feet. The average depth of the oceans is about 13,000 feet. The highest elevation of the land, Mt. Everest in the Himalayas, is 29,000 feet; the greatest known depth in the ocean, in the Pacific, is 35,400 feet. Thus the greatest difference in relief exceeds 64,000 feet, or more than 12 miles. Relative to its size,

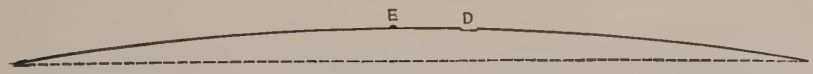

Fig. 4. — Mt. Everest in comparison to the size of the globe. Part of the arc of a globe with radius of one foot is shown; on this Mt. Everest, E, would be about $\frac{1}{50}$ of an inch high. D, in a similar way, shows the greatest depth of the ocean.

however, the globe has an extremely small relief, and it is therefore comparatively smooth; Fig. 4 shows its greatest roughness.

The relief features of the land are the *plains*, such as the Atlantic Coastal Plain; *plateaus*, such as that of the Colorado; and *mountains*, like the Appalachians extending from Canada to Alabama. In regard to the grouping of the relief forms of the Earth, certain facts are of interest and importance. The continents as a rule consist of basins that are bordered by mountain chains along the coastal rims, whereas the ocean basins generally have the reverse arrangement, the deeps being near the continents and the submarine ridges, or upswells of the bottom, being in mid-ocean. Some of the highest and most important ranges on the edges of the continents border the greatest deeps in the ocean floor, as for instance the Andes in South America and the partly submerged mountain chain that forms the Japanese islands and is the real eastern border of the continent of Asia; close to these ranges the ocean floor descends to great depths. It is not meant to imply, however, that mountains occur only at the continental edges, for they may extend in a wide zone far into the interior, as in western North America, or form systems crossing a continental mass, as in Asia.

Character of North America. — North America is regarded as the most typical of the continents. It is bordered by mountainous tracts on either side and contains the great basin of the Mississippi and its tributaries in the interior. The following broad features of the continent and especially of the United States will enter into many of the discussions of its geology.

On the east and south the continental shelf rises from the sea as the Atlantic Coastal Plain, and this plain extends to the base of a rugged mountainous tract of country, which stretches from Canada into Alabama and is known as the Appalachian Highlands; it includes the Appalachian Mountains. This mountainous belt gives way to the vast basin whose higher western part forms the Great Plains. The lower part of the basin is the Central Lowland, from which the Mississippi descends through the Gulf Coastal Plain. On the west the Great Plains abut upon the long series of north and south ranges that form the backbone of the continent and are grouped under the name of Rocky Mountain System. Between this system and the Pacific Mountain System, which makes the western rim of the continent, are the Intermontane Plateaus, consisting from north to south of the Columbia plateau, the Great Basin, and the Colorado plateaus. The Pacific Mountain System consists of the Sierra Nevada, Coast Ranges, and Cascade Range with the Pacific border lands. These relations and other minor ones can be seen in Fig. 5.

The divisions into which the United States and southern Canada are divided on the basis of physical features that have a common geologic history are shown on the map, Fig. 5. These *physiographic regions* are classified into major divisions, shown by letters, and minor provinces, indicated by the subscript numbers. Their names are given in the following list:

Major Divisions	*Provinces*
A. Laurentian Upland	A_1, Laurentian Plateau. A_2, Superior Upland.
B. Atlantic Plain	B_1, Continental Shelf (submerged). B_2, Coastal Plain.
C. Appalachian Highlands	C_1, Piedmont Province. C_2, Blue Ridge Province. C_3, Appalachian Valley and Ridge Province. C_4, St. Lawrence Valley. C_5, Appalachian Plateaus. C_6, New England Province. C_7, Adirondack Mountains.

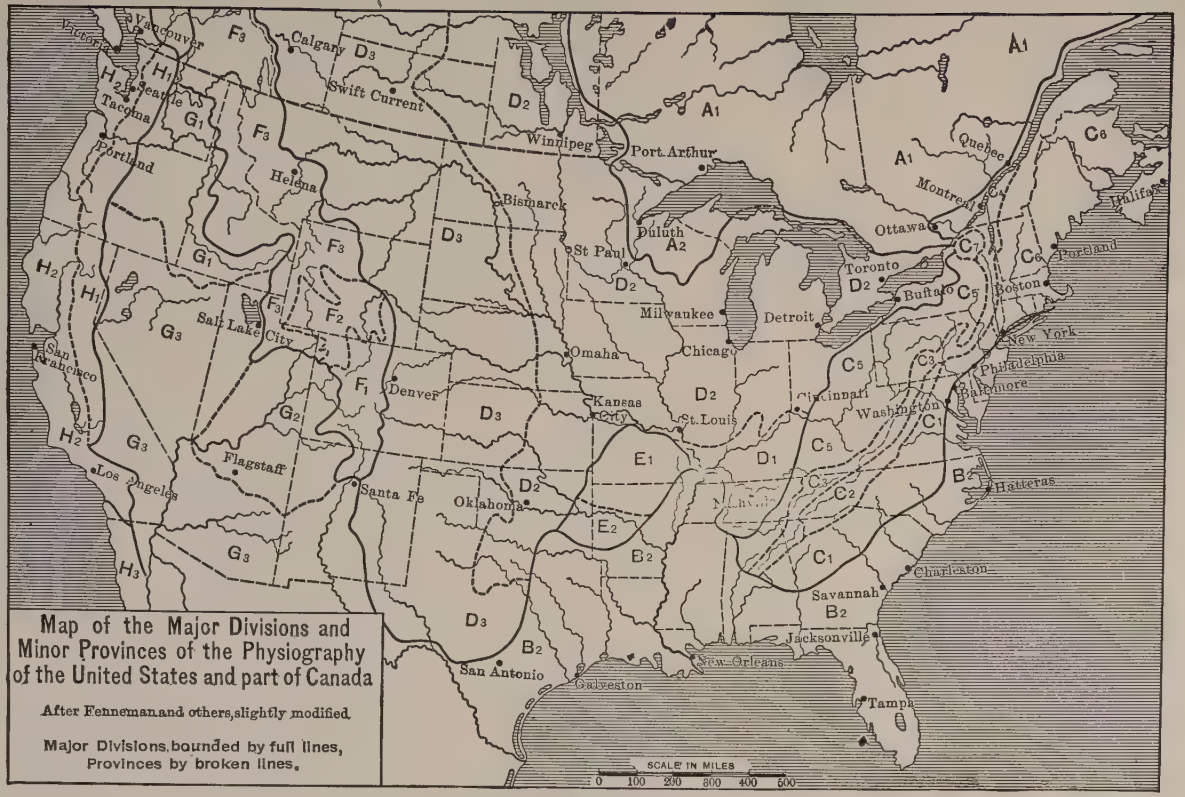

Fig. 5. — Chief physiographic divisions of the United States and part of Canada.

D. Interior Plains	D_1, Interior Low Plateaus. D_2, Central Lowland. D_3, High Plains.
E. Interior Highlands	E_1, Ozark Plateaus. E_2, Ouachita Province.
F. Rocky Mountain System	F_1, Southern Rocky Mountains. F_2, Wyoming Basin. F_3, Northern Rocky Mountains.
G. Intermontane Plateaus	G_1, Columbia Plateau. G_2, Colorado Plateaus. G_3, Basin and Range Province.
H. Pacific Mountain System	H_1, Sierra-Cascade Mountains. H_2, Pacific Border Province. H_3, Lower California Province.

THE OUTER ZONE OF THE EARTH; ROCKS

It is chiefly the outer zone of the Earth about which we have extensive and positive information. This outer zone is known as the crust; it is built of rocks, which are discontinuously covered by a thin mantle of soil. Because " crust " is thought by some to connote that the Earth has a liquid interior, the term lithosphere has been coined to avoid this implication, but we shall use the simpler term preferably. It is upon the crust that we live and exert our activities; we penetrate into it for fuels of various kinds, for metals, water, building material, and other mineral resources, all of which are essential to the physical side of modern civilization. A thorough knowledge of the component parts of the Earth's crust and its structure is consequently of the highest importance. The component parts of the crust are rocks, and we shall begin our inquiry by a study of the different kinds of rocks and the varied modes in which they occur.

Definition and Classification of Rocks. — The word rock, in the language of geology, means the material that composes one of the individual parts of the Earth's outer shell. According to their mode of origin, rocks are divided into three main groups: the *igneous* rocks, made by the solidification of molten material; the *sedimentary*, or bedded rocks, formed from sediments that were deposited chiefly by water (and to some extent by air and ice); and the *metamorphic* rocks, formed by certain processes acting within the Earth's crust on pre-existing rocks and partly or wholly destroying their original characters and producing new ones, so that the resultant rocks are best considered as constituting a separate group.

Thus we have three groups:

I. Igneous Rocks — consolidated molten masses.
II. Sedimentary Rocks — formed from sediments deposited by water, air, or ice.
III. Metamorphic Rocks — secondary, derived from preëxisting rocks.

Three-fourths of the land area of the globe is underlain by sedimentary rocks and the other fourth by igneous and metamorphic rocks. Although the sedimentary rocks thus preponderate in the visible part of the crust, they are essentially a veneer, a mile or less thick on the average. The foundation rock of the continents is largely igneous rock (granite), which is probably between 10 and 20 kilometers thick. The detailed discussion of the constitution of the crust is reserved for Chapter XV.

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CHAPTER III

THE ATMOSPHERE; WEATHERING AND SOILS

General Functions of the Atmosphere. — The atmosphere is directly essential to life on the Earth, and in addition it enables the sun and the rain to bring about life-giving changes on the Earth's surface. Without the atmosphere the Earth would be boiling hot by day or freezing cold by night, and devoid of life, as the moon is supposed to be. There would be neither wind nor rain nor bodies of fresh water. The lands would be rugged and desolate. Winding streams and soft landscapes mantled with colored soils and vegetation, and other vistas pleasing to the eye, would be lacking. For all these things, as we shall see, result directly from the presence of the atmosphere. Due to the rotation of the Earth and unequal heating by the sun, movements are set up in the atmosphere that account in part for our climate and weather. The wind itself builds up structures such as sand dunes, or abrades the rocky surface by the sand it carries along. But far more important are the chemical action of air and moisture, and the mechanical effects of changes in temperature, or frost action, aided by plant and animal life, that cause rocks to change into soil. This is called *weathering*, and the soil so formed may be swept away by the rain that falls from the atmosphere, and as more soil is formed from the rocks beneath, it in turn is carried off, and so the surface is gradually wasted away. This general process, involving the wearing away of the land, we call *erosion*. In this brief picture the importance of the atmosphere is evident, and we will now consider some of these factors in more detail.

Character of the Atmosphere. — The outer gaseous envelope of our globe is known to extend at least 200 miles above the Earth, since meteors heated to luminosity by friction with the outer air have been observed at this height. It must extend considerably higher. But even at a height of 50 miles it is extremely attenuated, and at 19,000 feet, the height of Mount St. Elias, its density is only half that at sea level. Its weight is estimated to be 1/1,200,000 of that of the Earth, or about 5 quadrillion tons. This weight exerts a pressure of nearly 15 pounds to the square inch, or a ton to the square foot, at sea level; but since gases transmit pressure equally in all directions, the pressure beneath any object is as great as that above, and hence the effect of the atmosphere's weight is not noticed.

The atmosphere consists of three chief constituents in the following quantities by weight:— nitrogen, about 75 per cent; oxygen, about 23 per cent; and argon, about 1.4 per cent. There are also present in small quantities the inert gases krypton, xenon, helium, and neon, but these are of no importance geologically. In addition, water vapor (average is 1.2 per cent at Earth's surface), carbon dioxide, hydrogen dioxide, ozone, ammonia, and traces of hydrogen, sulphur, organic matter, and suspended solids, are present in varying quantities. But of all these constituents the only ones that play an outstanding part in geologic processes are oxygen, water vapor, and carbon dioxide.

The oxygen, in addition to its obvious function of sustaining life, also enters into chemical changes that are important geologically. Wherever animals breathe or fire burns, oxygen is withdrawn from the air and locked up in compounds such as carbon dioxide. But growing plants, on the other hand, liberate some oxygen from carbon dioxide. Also the weathering of rocks involves the production of iron oxides such as limonite, in which oxygen is absorbed from the air. Thus the atmosphere is thought to be undergoing a slow net depletion of its oxygen content, though so slow that it is not detectable by chemical tests.

The carbon dioxide content in the atmosphere is relatively constant at 3 parts in 10,000 by volume. Though the quantity is small, its geological importance is great. Carbon dioxide is being added to the atmosphere from volcanic and mineral spring emanations, the combustion of fuels, the respiration of animals, and the decay of organic matter. It has been estimated that the consumption of coal annually returns to the atmosphere about 1/1000 of its present content, so that in 1000 years the amount of carbon dioxide in the atmosphere would be about doubled. But other operations are as continually diminishing the carbon dioxide content, notably its extraction from the air by growing plants, and its great consumption in the weathering of rocks.

The carbon dioxide and water vapor of the atmosphere also exert a beneficial effect on the Earth by causing part of the solar heat to be retained. Thus they help make the Earth hospitable to living things, and they aid chemical reactions that are important in the formation and the destruction of rocks. A comparatively slight increase in the carbon dioxide content of the atmosphere would bring a warmer climate, and a decrease of only a few per cent would lower temperatures appreciably.

CLIMATE AND WEATHER

It is obvious that rock weathering and erosion will proceed in a different manner in regions where rain falls intermittently throughout the year than in places where rainfall is seasonal; also the effect of these processes

will be unlike in an arid climate and a humid climate. Consequently, a knowledge of climate and weather is essential to a proper understanding of the character and distribution of the various climatic regions, and of the geologic processes that operate within them to produce far-reaching changes on the Earth's surface. And this involves consideration of atmospheric movements, moisture, temperature, and pressure.

Movements of the Atmosphere. — Air heated in the equatorial regions expands, rises, flows poleward, and settles down some distance north and south of the equator, and then moves over the surface toward the equator. The colder polar air tends to move equatorward. This circulation resembles that of the hot water in a house-heating system, and like the latter is induced and maintained by temperature differences. A great atmospheric circulation is thus set up. If the Earth did not rotate, and were smooth, and if its temperature varied uniformly from equator to poles, just such a simple atmospheric circulation would prevail. But this ideal simplicity does not exist. The rotation of the Earth from west to east throws this otherwise simple circulation very much askew, so that the air moving toward the poles is deflected eastward and that moving toward the equator is deflected westward. There are thus set up in the lower part of the atmosphere belts of *prevailing* or *planetary winds* (Fig. 6).

Those in the equatorial regions, roughly between latitude 30° N. and S., are the *easterlies* or familiar *trade winds*, and those in the middle latitudes are the *westerlies*. Where the trade winds from the northern and southern hemispheres meet near the equator, there is a belt of uprising air attended by cloudiness, high humidity, and calms, known as the *doldrums*. And between the trade winds and the westerlies there are the other belts of calms or light variable winds where the barometric pressure is high, the air descending, humidity low, and skies clear, known as the *horse latitudes*. Other movements due to differences in summer and winter heating over land and sea give rise to summer and winter *monsoons*, of so much importance to the climates of India, China, Australia, parts of Africa, and Texas. All of these are fundamental air movements that persist with regularity on different areas of the Earth's surface.

But in the middle latitudes there are, in addition, minor or secondary movements known as *cyclonic storms*, which affect animal and plant life from day to day and give rise to our weather. The westerlies of these latitudes are not simple prevailing winds always from the west. Here the equatorial and polar currents intermingle and, due in part to imperfectly understood effects of the upper atmosphere, and in part to unequal heating and cooling, circular air movements called *cyclones*

(low pressure areas) and *anticyclones* (high pressure areas) are set up. These are essentially thin, flat discs of moving air from 500 to 1000 miles across, that sweep easterly at an average velocity of 500 to 800 miles a day. Within these discs, the movement may be pictured as whirlwinds in which, in each of the low pressure areas (the *Lows*),

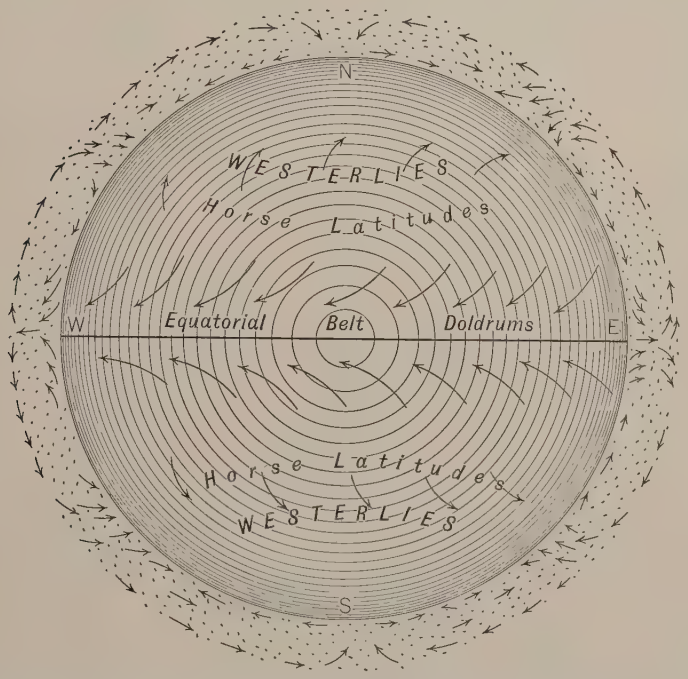

Fig. 6. — To show the direction of planetary winds on the surface and in the upper atmosphere. (Modified after Ferrel, and Tarr and Martin.)

a rising spiral of warm air rushes anticlockwise toward the center in response to the low pressure; whereas in each of the high pressure areas (the *Highs*) a descending column of cold air spirals toward the outside in a clockwise direction, being pushed out, as it were, by the high pressure. In the southern hemisphere the directions are reversed. About three Highs and three Lows may be expected to pass from west to east over any point in central and northeastern United States every two weeks. The direction of the wind at a given place at any particular time will depend upon which part of a Low or a High covers it; also, as the Low or High moves farther east, the direction of the wind at a given place will change. Thus the weather in the middle latitudes is continuously changing.

Atmospheric Moisture. — Water in the form of vapor is always present in the atmosphere, though its amount varies greatly from time to time and from place to place. When the atmosphere carries its full capacity it is said to be *saturated*, but this condition depends largely upon temperature, since warm air can contain more moisture than cold air. If air is only one-half saturated for a given temperature, it is said to have a *relative humidity* of 50. The average relative humidity over the oceans is about 85; over the lands it is considerably less. If air with a relative humidity of, say, 60 is cooled steadily, though there is no change in the actual amount of water present, the relative humidity will increase up to 100, or saturation, and further cooling will cause *condensation*. Upon this simple principle depends the control of precipitation, whether in the form of dew, cloud, fog, rain, or snow.

In nature there are several possible causes for cooling of moisture-laden air, such as radiation, moving of cooler air to it, or rising of the moist air into higher altitudes where expansion causes lower temperatures. The latter cause, depending on the simple principle of expansion, is the commonest. No surprise can be felt, therefore, at the heavy rainfall on mountains that lie in the path of air currents moving from the sea, for we know that high ground forces the air to move upward to cooler altitudes, where further cooling by expansion aids condensation into clouds and rain or snow. The same thing occurs when warm, moisture-laden air moves upward in the center of a low-pressure area, and that is why cloudy weather or precipitation accompanies Lows.

Conversely, if the air is warmed, its capacity to absorb and hold moisture is increased and it becomes undersaturated. Consequently, the cold air that falls in the center of a High becomes warmed by compression and is a drying wind from which no precipitation occurs. Similarly, air that falls on the lee side of a mountain range becomes warmed in its descent; if the fall be rapid and great, the skies are clear, the air is warmed, and any surface water in its path is rapidly absorbed. In the Rocky Mountains such a hot, drying wind is called a *Chinook;* in the Alps it is called the *Föhn*, and there the peasants welcome it because it melts the snows to water the pastures and it ripens the grapes and grain.

Atmospheric Temperatures. — The atmosphere receives most of its heat from the sun and, as pointed out above, this solar heat is retained largely through the action of carbon dioxide and moisture in the air. More heat is absorbed from the direct rays of the sun near the equator than from the slanting rays near the poles, and consequently the temperature, generally speaking, decreases from the equator to the poles.

But the uniformity of this arrangement is disturbed by ocean currents and the disposition of land and sea. Land absorbs more of the sun's heat than does water, and gives it off more readily; consequently it becomes warmer by day and cooler by night than does water. Water retains more of the heat it receives and tends to store it up until it is in part distributed by currents. For these reasons land climates are more variable than oceanic climates, and maritime regions are milder in winter and cooler in summer than continental areas. In the middle latitudes, local changes in temperature are sudden and pronounced, owing to the alternation of Highs and Lows.

Weather. — In middle latitudes the procession of Highs and Lows brings changing winds and temperature, and clear or rainy spells. Thus

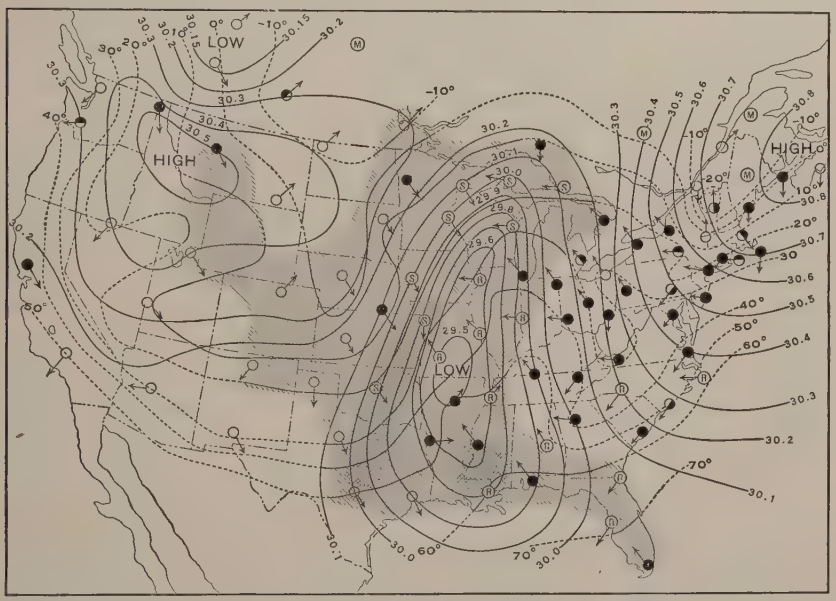

Fig. 7. — Weather map of the United States for Jan. 5, 1929. Shows high and low pressure areas, and a storm in the Mississippi Valley.

the weather moves and varies. In North America a low pressure area moving generally eastward may in its broad reach cover a considerable part of the eastern United States and southern Canada (Fig. 7). Rains occur on the eastern and southern side of the Low because the winds are warm and moisture-laden, after their sweep over sea waters to the south and east. When they ascend in the cyclonic center, cooling takes place and condensation soon occurs, so that cloudy or rainy weather accompanies the Lows. This eastward movement of Lows, then, gives

a generous rainfall to the eastern part of the continent. Occasionally the Low may remain nearly stationary, and then rains may be so continuous that floods result.

As a Low passes and a High follows, the warm winds from the southeast and south shift to the west and northwest and the temperature falls rapidly. This causes the cool spells of summer and cold waves of winter. The descending cool air is undersaturated; drying winds and clear skies result.

Meteorological observations gathered from scattered points make it possible for weather maps such as Fig. 7 to be plotted daily, and by use of the principles just discussed and with the understanding that weather travels, reliable weather forecasts can be made 24 to 36 hours in advance.

Climate. — Climate has been defined as the "sum total of meteorological conditions that constitute the average state of the atmosphere at a given point on the Earth's surface."

If the Earth had a smooth, homogeneous surface and no atmosphere, there would be simple solar climates whose distribution would correspond with latitudes. While it is true that latitude is the most important single factor controlling climate in so far as temperature is concerned, there are other factors that profoundly modify it. Some of these have already been mentioned, such as proximity to large bodies of water or ocean currents, prevailing or seasonal winds, mountain ranges, cyclonic storms, and altitude.

Away from the equatorial belt the ocean currents are the chief controlling factor in the climates of lands that border the seas. For example, compare the equable climate of the British Isles, tempered by winds from the warm Gulf Stream, with that of Labrador at the same latitude, washed by the chilly Labrador current. Or compare the mild climate of the coast of British Columbia and Alaska, where the influence of the Japan current is felt, with the harsh climate of the Siberian coast directly opposite. This contrast between east and west coasts is present nearly everywhere. A region in which the winds blow from an adjacent ocean has, for its latitude, mild winters and cool summers and small daily variation in temperature, features typical of a maritime climate. But it will be noted that in high latitudes true maritime climates are restricted to the west coasts of continents, whereas the east coasts, at corresponding latitudes, are influenced by continental conditions.

In the regions of the monsoons, notably in Asia, the year is clearly divided into a dry season and a period of torrential rains, when moisture-laden winds blow toward the lands. If these winds are intercepted by mountains, there is a particularly heavy rainfall for the reasons previously discussed.

Continental climates are marked by great variation in temperature, rainfall, and barometric pressure. In the United States, for example, a thin maritime belt (Fig. 8) fringes the northwest coast where extremes in temperature are few and rainfall is plentiful. The prevailing westerly winds carrying moisture from the Pacific Ocean are forced high by the coastal ranges, whose western slopes are thus well watered. Where the dry winds drop suddenly on the eastern lee slopes, there are arid and semiarid belts where temperature changes from night to day are pronounced. The contrast with the coastal belt is striking. Throughout

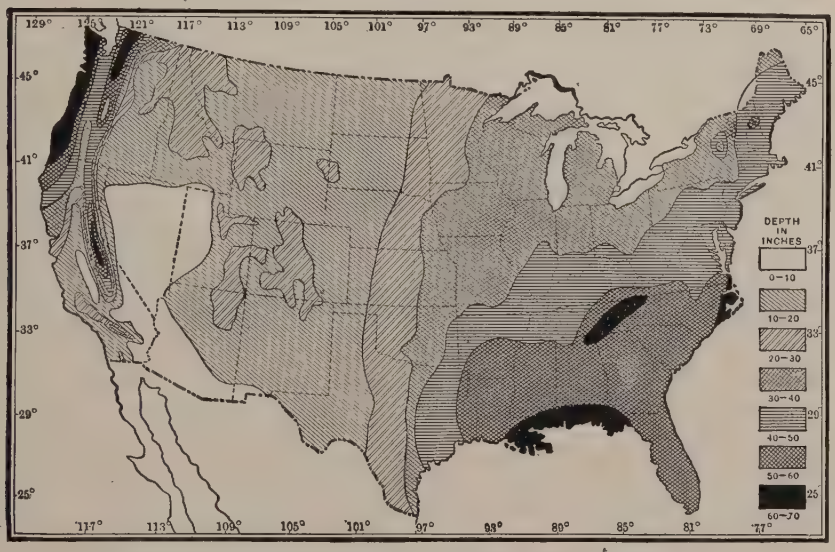

Fig. 8. — Rainfall map of the United States. (U. S. Geol. Surv.)

the Great Basin region arid or semiarid climates prevail and the daily and annual variation in temperature is extreme. The climate of the Mississippi drainage basin and the Eastern States is controlled largely by latitude, cyclonic storms, and proximity to the Atlantic Ocean. Sudden changes in temperature are frequent, and clear and rainy weather alternate throughout the year. The New England States, over which most of the cyclonic storm paths converge, have particularly changeable weather, with fairly evenly distributed precipitation throughout the year, and pronounced temperature changes. Warm and cold spells alternate with surprising suddenness, even in midwinter, due to a High following a Low. This is typical of a continental climate in middle latitudes. An extreme example of the continental climate is central Asia, which is not only remote from the oceans, but is also cut off from

ocean winds by mountain barriers. In consequence of these various factors, different climatic provinces are produced throughout the world.

Factors that Change Climates. — From what has been said above it is clear that if the controls of climate be changed, then a different climate for a given place will result. Examples are changes in the distribution of land and water, with a shifting of the ocean currents; the formation of a new mountain range on the land; changes in solar radiation, or other astronomic changes; and changes in the carbon dioxide content of the atmosphere, or in the atmospheric circulation. The studies in geology to follow will teach us that in the past just such changes of climatic controls have taken place.

Climates of the Past. — The records of geology show that there have been profound changes of climate in the course of the Earth's history, and such changes are probably still in progress. Fruitful lands of today were once barren deserts; corals and subtropical plants thrived in what are now cold northern regions. And in the period just preceding our own, a widespread glacial climate prevailed over the regions of high latitudes, and ice sheets spread over the lands. Still earlier, glaciation spread over subtropical parts of Africa, Australia and India. Even within historic times changes in increased rainfall or in progressive desiccation have been noted. An example of the latter is to be seen in the part of northern Africa that was once the granary of Rome.

These climates of the past, in their changing character throughout the ages, have influenced notably the development of life upon the Earth, and have affected rock weathering and other geologic processes that have operated within their confines.

WEATHERING

The records of geology teach us that the Earth is constantly changing, and nowhere is this so evident as on the Earth's surface where the atmosphere and the lithosphere meet. Even the hardest rock cannot withstand the slow unrelenting attack of the atmosphere; it will crumble away to fragments or soil.

The attack of the atmosphere upon the rocks is twofold — mechanical and chemical. The first produces *disintegration* of the rocks by mechanical agencies such as frost action, changes in temperature, wind or rain. The chemical attack results in *decomposition* of the rocks, and a certain amount of this is essential to form good soil. Both kinds of destruction usually operate together and each is aided somewhat by the work of plants and animals. The dominance of one or the other depends to a large degree upon the nature of the climatic provinces in

which they operate. Together they constitute a rather complex set of processes called *weathering.*

Weathering is aided greatly by the shattered character of the bedrock, which is penetrated in all directions by cracks and fissures, some great, some small (Fig. 9). Even the mineral grains of the rocks are more or less filled with cracks. These openings are potent factors in promoting

Fig. 9. — Shattered nature of the bedrock which facilitates disintegration and decomposition by allowing ready ingress of water, air, and plant roots.

disintegration and decomposition, for they allow ready entrance of air and water and plant roots into the rock.

What these various agencies are that together produce weathering it will now be our purpose to consider.

Mechanical Agencies

The Expansive Force of Freezing Water. — Water pipes that burst upon freezing are a familiar manifestation of the expansive force of freezing water — a force that exceeds 2000 pounds per square inch In the days of muzzle-loading cannon, captured enemy guns were frequently disposed of by being filled with water and left to freeze, by which means they were split from end to end. The effect upon rocks is similar; water trickles into the numerous crevices, joints, and pores (Fig. 9), and upon freezing, pries off small pieces or large blocks. The latter are in turn similarly broken into smaller pieces. In the same manner porous sandstones may be completely reduced to sand. This action goes on whenever water freezes in a confined space. Naturally it is most pronounced where freezing and thawing occur frequently, as in most

temperate regions in the fall and spring. This is particularly the case on high mountains where, almost regardless of latitude, thawing by day and freezing by night may go on over a considerable part of the year. Anyone who has camped on a steep mountain on a frosty evening will never forget the startling noise of falling rocks dislodged by this process. Débris formed in this way contributes to the formation of *talus slopes* (Figs. 10, 14), containing many millions of tons of broken rock, which flank the lower mountain slopes, particularly in high latitudes. Those who climb mountains are familiar also with the masses of broken rock

Fig. 10. — Rock disintegration and weathering in high altitudes with formation of long talus slopes of slide rock. Mt. Sneffels, Colorado. (U. S. Geol. Surv.)

on the tops of ridges. These are impressive illustrations of the effectiveness of frost action on larger topographic features. The results of this process are also clearly evident on many buildings and tombstones throughout the northern States, and should be taken into account in the proper selection of building stones. Had the Sphinx been set up in New England instead of in the warm, dry climate of Egypt, its face would not be recognizable today.

Frost action, even by itself, is an important factor in mountains and in temperate and cold climates, in helping wear away projecting rock masses; the accomplishment of a single year may be small, but in long ages the accumulated results are of great magnitude. Further, it provides smaller fragments upon which chemical attack may be made more effectively. But it must not be assumed that all heaps of broken " slide rock " are caused only by frost action, for several other processes of rock disintegration produce similar results.

Changes in Temperature. — Most substances when heated expand, and if they are good conductors of heat, like metals, the expansion ex-

tends well beyond the actual place of heat application. An iron rod cannot be held with the free end in the fire without discomfort to the hand, but a piece of rock can be so held. It is a poor conductor of heat. If the heated end expands and the rest does not, the two must separate. When cooling occurs an outside layer that contracts becomes too small for the unchanged interior; it cracks. A similar process goes on, but much more slowly, when rocks are heated by the sun and contract in the shade (Fig. 11). On upstanding rock masses the projecting corners

Fig. 11. — Buckling in sandstone layers due to expansion from heating by the sun. Wyoming. (U. S. Geol. Surv.)

are spalled off first, then thin shells peel off or *exfoliate*, like onion layers, as illustrated in Fig. 12. Boulders become spherical and gradually disappear. Larger masses take on a rounded form, like some dome-shaped hills. The scales that drop off become further broken up and dust-like material results. Similar effects, produced in a much shorter time, are seen in a stone building that has passed through fire. Dr. Livingstone found in Africa that rock surfaces heated to 137° by day, upon cooling rapidly at night threw off with sharp reports angular fragments up to 200 pounds in weight; and according to Stanley, contraction caused by cold rain falling on those sun-heated African rocks causes them to split and exfoliate. Still another though minor effect is produced by change of temperature; unlike mineral grains composing a rock neither absorb heat nor expand equally; consequently minute interior strains are set up that eventually disintegrate a surface layer of the rock.

Thus *disintegration* proceeds, and broken rock results. Rain or wind may carry away the disintegrated material and leave fresh rock surfaces

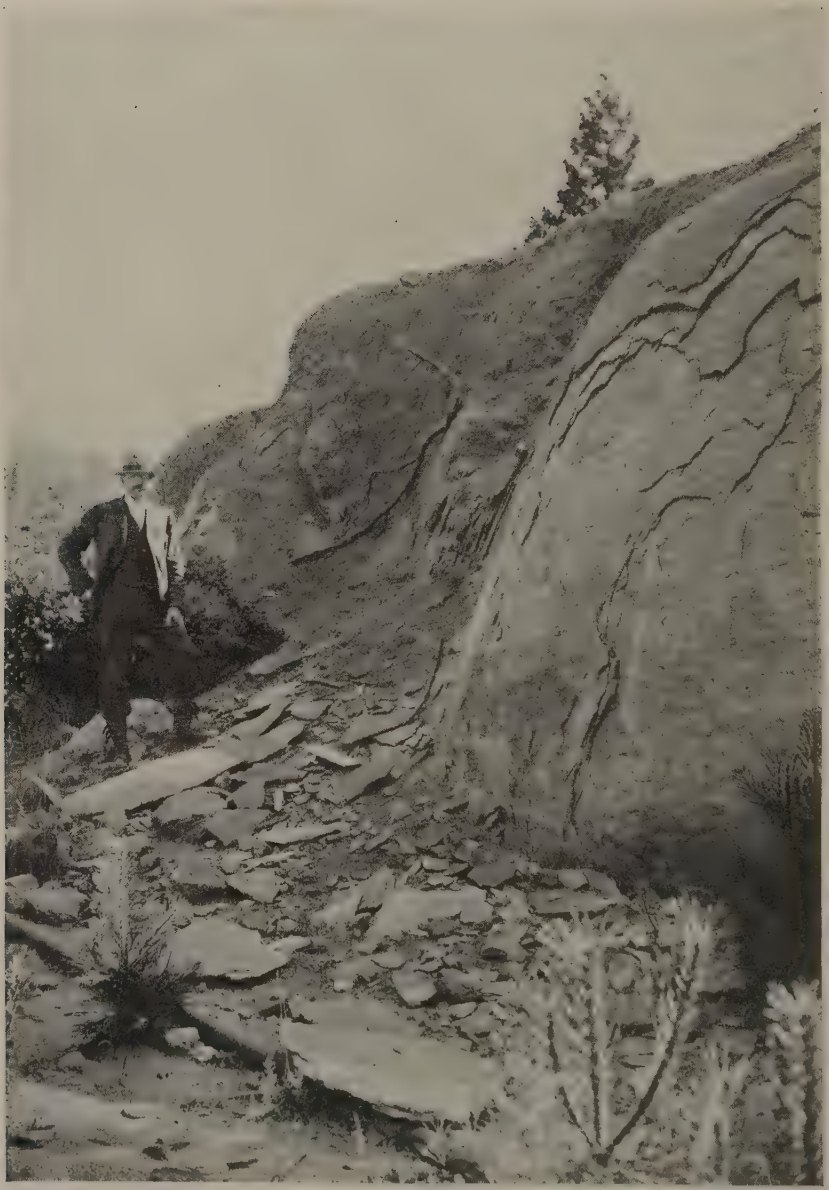

Fig. 12. — Exfoliation, or scaling of rock, by alternate expansion and contraction of surface layers. Nevada City, Cal. (U. S. Geol. Surv.)

exposed to further attack. And so the process repeats itself until exposed rock surfaces finally disappear. Hills or cliffs are thus worn down, or jagged valley slopes are smoothed into gentle curves.

The process is most rapid in regions where there is a pronounced difference in temperature between day and night or between winter and summer. In temperate climates the difference between the heat of summer and the cold of winter may be 100° or even 150°. And in desert climates the daily range may be 50° or exceptionally 100°. Therefore high latitudes and altitudes or semiarid regions favor disintegration by temperature changes more than do regions with a moist, equable climate.

Rain. — As compared to the agencies previously discussed the mechanical work of rain is small, although it is an important agent of chemical weathering and of erosion. The impact of rain on steep slopes composed of soft rocks tends to dislodge small particles which are readily washed or blown away. New surfaces are thus exposed to a repetition of this process, which in arid regions helps to form " badland " topography. The mechanical work of rain is also seen in the softening of rocks or clays, enabling them to be more readily removed by rain wash or streams.

Wind, when it causes sand blasts, is also a mechanical agent of abrasion, but its work is of such broad scope that it is treated separately in later pages.

Plants and Animals. — The wedging apart of rocks by the roots and trunks of growing trees is a familiar sight to the observer of nature. In their slow expansion by growth they exert a powerful disruptive force which even large masses of rock are unable to withstand. Likewise, the rootlets of plants and shrubs insinuate themselves into little crevices of bedrock and boulders, and slowly break them apart. In every little crack where seeds may lodge and grow this process goes on. In the course of time the amount of work done in this way must be great, but its chief importance is in helping chemical attack to proceed more readily.

Animals such as moles, gophers, worms, and ants, by making holes and burrows in the soil, aid disintegration to a slight degree. Their work enables the other agencies to come more readily into contact with covered rock surfaces. The most destructive animal, however, is Man. He has felled the forests and otherwise destroyed the protective covering of plant life over the soil, allowing it to be carried away by rain or wind and thus expose fresh surfaces of rock to processes of disintegration.

Chemical Agencies

Chemical agencies, like mechanical agencies, are essentially superficial in their work. They act only on a thin skin of the outer rocky crust, and lose their effectiveness at shallow depths. Their work is inextricably interwoven with that of the mechanical agencies, and it is only for convenience of discussion that we consider them separately. The result of their combined attack is the soils from which we derive our sustenance.

Decomposition or decay may act directly upon solid bedrock, but it proceeds much more readily upon rocks previously fragmented by disintegration processes. It goes on more slowly than disintegration, and is favored by moisture and heat, and retarded by cold. Consequently, unlike disintegration, it is promoted by warm, moist climates, by moderate relief, and by a covering of vegetation. If the climate of Egypt were less dry, fewer pieces of old stone art would have been preserved for us; and if the climate of Labrador were less cold, a productive soil mantle might exist there.

The chief agencies that bring about decomposition of the rocks are oxygen and carbon dioxide from the air, water, the organic acids released by decaying vegetation, and certain bacteria.

Decomposition. — If a piece of old iron is left exposed in a damp climate, it rusts; that is, it changes from iron, which is unstable in the presence of oxygen and water, and entering into combination with these substances, forms rust or limonite (iron, oxygen, and water), which is stable under the changed conditions. A simple chemical reaction has taken place, and a new substance has resulted. This is one of the changes that occur when rocks decay. The iron-bearing minerals of the rocks, for example, yield limonite, and the rock surface or resulting soil takes on the familiar tints of iron oxide, imparting to deeply weathered areas the red, yellow, or brown colors so pleasing to the eye.

But to produce fertile soil, other rock minerals must be decomposed. The chief rock-making minerals are the feldspars, and they are also important soil-yielding minerals. The chemical change of one of them, orthoclase, may be illustrated: —

Orthoclase + Water + Carbon dioxide yields Clay + Silica + Potassium carbonate

$[KAlSi_3O_8 + H_2O + CO_2 = (\text{Various hydrous aluminum silicates}) + SiO_2 + K_2CO_3]$

This is one of the most important reactions that take place in nature, since the existence of life so largely depends upon it. The necessary carbon dioxide is obtained from the atmosphere and in part from decaying vegetation, and enters readily into solution with water to form car-

bonic acid. Clay is an essential ingredient of good soils, and potassium carbonate is a necessary food of plant life. The potash chemically locked up in orthoclase is thus set free as a soluble food that can be assimilated by plants.

Similarly, other common rock-making minerals that contain aluminum, such as amphibole, chlorite, and mica, yield clay. Some also yield potash. Quartz, being relatively insoluble, resists decomposition and remains in the soft decomposed materials as grains of quartz, or sand. When granite, composed chiefly of feldspar and quartz, is acted upon, the resulting soil is a mixture of clay and sand grains, called loam.

The combination of a substance with oxygen is *oxidation;* with water, *hydration;* and with carbon dioxide, *carbonatization;* all these processes enter into decomposition. As a result new substances called oxides, hydrates, or carbonates, are formed. Some remain behind in the soil, others may be carried off in solution. Hydration is usually accompanied by an increase in volume, and this swelling is an additional factor in helping to break up the rocks. Some geologists think that this effect of hydration is very important in producing some of the effects of disintegration ascribed to change in temperature.

Solution. — In addition to the more complex chemical changes mentioned above, simple solution also aids chemical decomposition. Pure water (H_2O) is a poor solvent for rock minerals, but when it is combined with carbon dioxide (CO_2) to form carbonic acid (H_2CO_3) it is a powerful natural solvent. It attacks, among other substances, calcium carbonate ($CaCO_3$) and converts it into calcium bicarbonate [$H_2Ca(CO_3)_2$], which is quite soluble in water. Some rocks, such as limestone, are composed almost entirely of calcium carbonate, which forms the mineral calcite, and in others, such as sandstones, this substance acts as a cement to bind the individual sand grains together. When carbonic acid acts on such a rock, the binder is dissolved, the grains loosen, and the rock crumbles and breaks down into sandy soil.

In the case of limestone, the greater part of the rock may be slowly dissolved, leaving behind only the insoluble impurities, usually clay, to form a *residual* soil. Unbelievable thicknesses of rock have been removed by this process, at a rate that in places may be as great as 1 inch in 25 years. The impurities gradually accumulate on the surface, giving rise to the fertile soils of such limestone regions as southern Kentucky.

The impurities in the limestone may, in places, consist of iron or manganese compounds instead of clay, and residual accumulations of these substances have formed valuable deposits of iron or manganese ore that are being mined in different parts of the world.

Solution is also effective in removing some of the soluble products that result from chemical decay.

Plants and Animals. — Both plants and animals aid in the chemical attack upon rocks. Plants in their growth extract carbon dioxide from the atmosphere; they release the oxygen and retain the carbon. On their death and decay some carbon dioxide is released and furnishes a source of carbonic acid for chemical attack upon the underlying rocks. As the plants die and others grow upon their ruins, a part of the carbon may accumulate as a carbonaceous residue, known as *humus*, desirable as an ingredient of fertile soils.

The humus also supplies to the rain water that seeps through it small quantities of complex organic acids which are in themselves effective chemical agents of rock destruction. They also supplement the work of carbonic acid in mineral decomposition or solution.

In addition, these acids are potent reducing agents; that is, they are able to take away oxygen from some oxidized compounds. This is most strikingly seen in the effect upon the ferric hydroxides that color the soils red, brown, or yellow. The ferric hydroxides (Fe_2O_3 + water) are reduced to ferrous oxide (FeO), some of which unites with carbon dioxide to make ferrous carbonate, and this in turn is soluble in carbonic acid. Thus some of the iron is leached out and removed, and the remaining part no longer gives high colors to the rocks or soils. This is strikingly seen where red sandstones overlain by vegetable mold are grayish in hue just beneath; and where rootlets extend down into red rocks there is a bleached zone surrounding them. Where ferric oxides, limonite or hematite, form the cement that binds the grains of sedimentary rocks together, its removal by the action of organic acids causes the rock to fall to pieces and so brings about the formation of soil. Where much organic matter is present in soils, vivid colors are absent. The two are incompatible. For example, in certain clays dark blue, dark gray, or greenish to black color denotes the presence of organic matter; and any iron present must be in the colorless ferrous form. But if these same clays are fired to make bricks, the organic matter is burned out, the ferrous iron is oxidized to the ferric state, and red brick results. If iron is not present, white brick is formed.

Insects and other animals that live and move about in the soil aid chemical work by upturning the soil and exposing fresh surfaces to weathering. Their openings in the soil also enable weathering agents to reach more readily the underlying bedrock. Thus, Darwin states that in England earthworms bring to the surface 10 tons of mold to the acre every year, and Branner believes that in many tropical regions ants are even more effective in upturning the soil.

Bacteria also contribute their part. Certain forms have the unusual power of assimilating carbonate of ammonium and setting free nitric acid which attacks and decomposes rock minerals. They penetrate in great numbers every little nook and cranny in the soil or bare rock, and in time their effect is of no inconsiderable geological significance.

The general effect of chemical decomposition is to break up the complex rock minerals to form softer incoherent materials that may remain as a soil mantle, or may readily be carried away by rain or wind.

RESULTS OF WEATHERING

The results of weathering are no less numerous than are the physical and chemical agencies that bring them about. It is unusual, as we have seen, for one alone of these agencies to prevail for any time; they operate together. The sum total of their work destroys the integrity of the surface rocks, fashions the smaller details of the landscape, and spreads a life-giving soil mantle over most parts of the earth. By means of them some of the rugged details of mountain scenery have been sculptured and the gentle slopes of softer vistas have been smoothed. Materials are supplied to build flood plains and deltas and sedimentary rocks. Some of these results of the coöperative work of the agencies of weathering may now be considered, and of these, soils stand first.

Soil and Rock Mantle. — Nearly everywhere the Earth's surface is mantled by a thin veneer of soil or of disintegrated rock. Its thickness varies considerably; compared with the Earth as a whole, it is but a film. In tropical regions it may extend a few hundred feet in depth. Normally it is only a few feet, or a few tens of feet in thickness. Soil is formed, as we have seen, chiefly from the breaking up of rocks; decayed vegetation in most places adds to it but little. In the agricultural sense the term soil is applied only to that upper portion or *topsoil* which contains some humus and supports plant life, but in the geologic sense it embraces as well the underlying rotted rock or *subsoil*.

Soils may be divided into two large groups: *residual soils*, which have been formed in place from the immediately underlying rock; and *transported soils*, which have been moved from their place of origin.

Residual soils normally pass gradually downward from a topsoil supporting vegetation into the subsoil of coarser material full of bits of rotted rock, and then imperceptibly into decayed rock that crumbles more or less easily, and this in turn merges into unaltered solid bedrock (Fig. 13). This gradual transition from topsoil above to solid rock below is one proof that the soil has been formed in place by the decomposition of the underlying rock. The transition between rock and soil is not

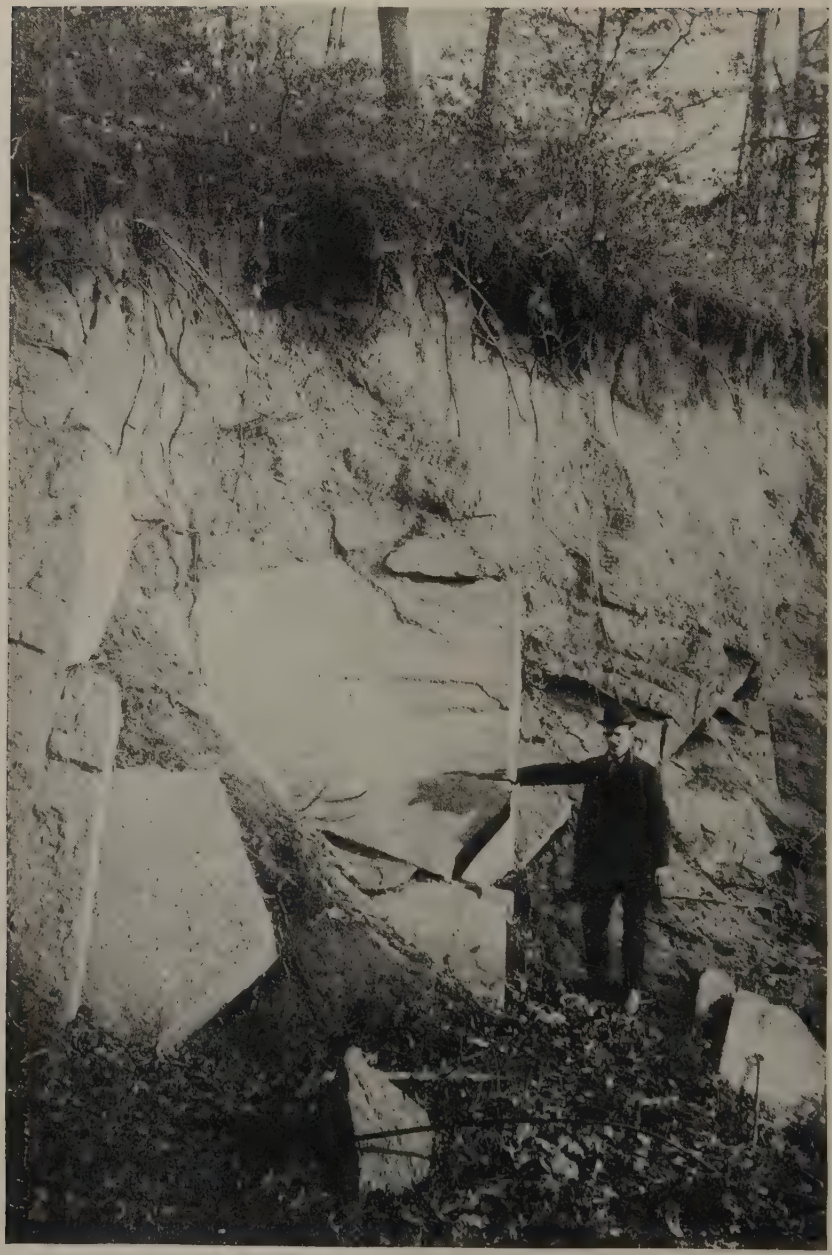

Fig. 13. — Residual soil formed by rock weathering and decay. The material graduates from firm rock below, through rotten rock and then subsoil, to true soil above. The transition is gradual. The soil above is colored dark by decayed organic matter. (Merrill, U. S. Nat. Mus.)

only gradual but also highly irregular from place to place. Rich residual soils are widespread over most of the Piedmont district of the southeastern United States, and are the chief soils in humid tropical or subtropical regions.

Residual soils in general are composed of much finer materials than transported soils, for in them decay has been more complete.

Transported soils have been shifted by rain or running water (*alluvial*), by wind (*eolian*), by moving ice sheets (*glacial*), or by gravity down hill slopes (*colluvial*). Consequently they vary in composition more than residual soils and range from the finest silts to gravel. They have no essential kinship with the underlying rock and their source may have been far distant from their present site. Wide areas of the northeastern and north central United States are mantled by glacial soils that came from farther north, and the fertile silts that flank the lower Mississippi River came from far upstream.

Character of Soils. — The character of soils, whether they are residual or transported, depends largely upon the kind of rock from which they were derived and upon the climatic conditions under which weathering took place. Thus rocks composed chiefly of feldspar under arid conditions disintegrate to feldspathic sand; under humid conditions they yield clay. A pure sandstone yields only sand. According to the size of the particles which compose the rock mantle, the following gradations are recognized: Pieces of loose rock from the size of a small melon up are termed *boulders;* those larger than peas are called *gravel;* pieces smaller than peas, but not coherent when wet, are *sand;* and the finest material, which can be carried by the wind, is *dust;* the last is termed *silt* or *clay,* according to its character, and generally coheres when wet. Ordinary soils are composed of variable mixtures of sand and these finer materials.

Loam, a mixture of clay and sand, is easily worked and makes excellent soil; pure clay contains abundant plant food but is apt to be stiff and difficult to work; very sandy soils are usually unfertile.

Soils very rich in humus are called *muck.* When a clay soil contains a considerable quantity of calcium carbonate it is termed a *marl.* Thus *sands, loams, clays, mucks,* and *marls* are the chief kinds of soils and there are all gradations of these into one another. Owing to the presence of the dark organic matter or to the greater oxidation of the iron compounds, and often for other reasons, the topsoil is likely to be much more strongly colored than the underlying subsoil.

Talus. — On mountains, where the prying action of frost, or exfoliation by changes of temperature, or swelling by hydration take place, the rock so broken collects by gravity on the lower slopes, forming *slide rock* or *talus.* This is illustrated in Figs. 10, 14. The talus

may be a sheet-like form that flanks the mountain slope, as in Fig. 10, or it may be cone-shaped, when it is called a *talus cone*, as in Fig. 14. A mountain may become so buried in its own talus that only its summit projects. Those who have climbed mountains realize the abundance and size of talus slopes, particularly in high latitudes. They are, however, but transient features, for the materials composing them eventually are comminuted and are largely carried away by water or by wind.

Fig. 14. — Mt. Washington, Oregon, showing talus cones. (U. S. Army Air Corps.)

Residual Boulders. — The change of bedrock into soil is apt to proceed first along cracks and fissures. These usually intersect each other and inclose large and small blocks of bedrock. Consequently the agents of decomposition and disintegration begin their work on the outside of such blocks. The vulnerable corners succumb, then the sides are attacked; rounded *residual boulders* may result. Where the attack of chemical and mechanical agencies has gone so far that rock masses or boulders take the shape of spheroids, it is called *spheroidal weathering*.

Concentric layers, like the skins of an onion, may continue to peel off until eventually the rock bodies disappear. Also, masses distributed through the bedrock may be different in composition or texture from the rest, and thus harder or less soluble. These may also be left as residual boulders (Fig. 15). Some residual masses may rest in apparently

Fig. 15. — Residual boulders left by decomposition and wearing away of bedrock. The boulders are included masses of a harder, more resistant material and of rounded shapes (concretions). This shows that some residual boulders differ from the bedrock on which they lie. Coalinga, Cal. (U. S. Geol. Surv.)

unstable positions (Fig. 16). It must not be assumed, however, that all boulders are formed by this process, for many are transported blocks whose composition is unlike the underlying rock, and many have been formed by differential disintegration or wind erosion.

Exfoliated Forms. — Exfoliation has also operated on a grand scale, not to be compared with the thinner veneers of disintegration by spheroidal weathering, to produce great picturesque domes such as those of the Yosemite Park and Stone Mountain, Georgia (Fig. 303). On Stone Mountain huge slabs, scores of feet across and up to a foot in thickness, are in various stages of peeling off; doubtless hundreds of feet have in the past been removed from the top and sides of the mountain in this manner. Chemical decomposition seems to have played little

part in producing this result, since the loosened material, though friable, is quite fresh. Changes in temperature must have been the dominant agent of destruction, though frost action and hydration may have played a minor rôle.

Fig. 16. — Residual boulder resting on bedrock. "Balanced Rock," Garden of the Gods, Colorado.

Differential Weathering. — Nature has tooled rocks of different resistance to erosion into grotesque forms and scenic bits of landscape such, for example, as may be seen along the Cody entrance to Yellowstone Park. Softer or more soluble parts have been removed and the harder resistant parts are left to stand out in relief (Fig. 107). They may be only small ribs of insoluble quartz that project from the surface of soluble limestone, or "mushroom rocks" or pagoda-shaped boulders. Or they may be larger features such as balanced rocks (Fig. 16) and pinnacles or columns, or even a succession of cliffs and benches where hard and soft layers of rock alternate, such as may be seen on a grand scale in the Grand Canyon of Arizona. The more common balanced rocks, pinnacles, and pedestals, such as attract the visitor in the Garden of the Gods or Monument Park, in Colorado, are formed usually by the more rapid weathering, chiefly by mechanical agencies, of softer beds of rock that underlie harder strata. The caps of harder rock eventually fall to the ground and, until they are disintegrated, remain as boulders.

WORK OF THE WIND

The geologic work performed directly by the wind is mechanical in its nature; it is a builder as well as a destroyer. It carries away products of weathering; it sweeps sand against rocks and abrades them; and it builds up dunes and dust deposits. Its work is most effective where vegetation is absent; consequently its results are to be seen in arid or semiarid lands and near sandy shores. The effect of the wind in forming waves that waste the coasts may best be deferred to the discussion of oceans.

The Wind by Itself. — The wind by itself plays a minor rôle in blowing away the products of weathering as they are formed, thus exposing fresh rock surfaces to further wasting. With its aid, therefore, disintegration and decomposition can proceed faster. This action proceeds best on hillsides or valley slopes where lodgment of weathered particles is insecure and where gravity aids; but it also operates even in flattish areas, particularly in arid regions, where mechanical agencies disintegrate the surface, and wind blows the material away, making shallow depressions. Such wind work is termed *deflation*. In semiarid regions such as Nebraska, in times of drought, the wind dries the soil and blows it away from cultivated fields, with disastrous results to young crops and the supply of soil.

The wind also transports vast quantities of material. A gentle breeze lifts and carries dust, a strong wind drives sand (Fig. 17), and a tempest

Fig. 17. — Sand storm sweeping over Khartoum North; in front is the Blue Nile. Shows enormous transporting power of the wind. Soudan, June 6, 1906. (Wm. Beam.)

can move gravel the size of peas. For example, a single storm that travelled from the arid Southwest a thousand miles to the Great Lakes region brought with it a million tons of dust, which was deposited on the snow over a wide area. Such material, dropped when the wind slackens, may eventually form deposits of great magnitude. They are known as *eolian* (Aeolus, god of the winds) deposits, to distinguish them from water-laid deposits.

Abrasion by Wind-blown Sand. — The smoothing effect of sandpaper applied to a rough surface is a familiar example of abrasion. The

particles of sand fastened to the paper make sharp and effective cutting tools. Similarly, begrimed stone or brick buildings today are being renovated by a sandblast — sand blown by compressed air. Nature supplies a similar agent of abrasion in the sand moved by wind. Though less efficient as a cutting tool than its artificial counterpart, it is, over long periods of time, an important agent of rock destruction.

The rate of cutting by storm-driven sand may be quite rapid. For example, window panes of houses along seashores have lost their transparency in a day or two, and have been completely penetrated in a

Fig. 18. — Rounded rock forms produced in part by wind erosion. Note irregularities in surface. Arizona. (Bateman.)

month or so. Wooden telegraph poles planted in the desert are in some localities quickly cut down by the sand drifting past their bases, and on stretches of railroad that cross deserts, rails have to be replaced frequently because of abrasion by drifting sand.

The most effective abrasion by wind-driven sand is to be seen in arid regions and deserts, where rainfall is scanty and protective vegetation is sparse. Exposed rock masses are carved and sculptured into rounded (Fig. 18) and " hoodoo " forms. In detail, the surfaces are commonly pock-marked or intricately *fretted* by the removal of softer parts. Further evidence of eolian abrasion is to be seen in the smoothed, polished pebbles that lie in such areas. Many of them are faceted into angular forms, commonly with triangular cross sections. Such pebbles are called *dreikanter*.

Sand Dunes. — Sand dunes are hillocks made by wind-borne sand in a manner similar to that in which snow forms drifts. Their growth is started by some obstacle, such as a rock, a bush, or a surface irregularity, that gives local protection from the wind or causes wind eddies, so that the moving sand is deposited. A dune, once started, continues to grow and may reach a height of 100 feet to 200 feet or more. Even the erection of buildings starts their formation in some places. The surface of a dune is commonly covered with fine parallel ridges of sand an inch or so in height, transverse to the direction of the prevailing wind, and called *ripple marks* because they resemble the markings on sand made by water ripples (Fig. 19). Dunes are formed wherever free-moving

Fig. 19. — Sand dunes near Mammoth Station, Cal., showing ripple marks. (U. S. Geol. Surv.)

sand exists and where there are prevailing winds. Frequent changes in wind, no less than vegetation or exceptionally rough topography, are unfavorable to their development. They are found in all parts of the world along low coasts where sand made by waves is washed ashore and, caught up by the prevailing winds, is drifted inland and accumulated. Dunes so formed occur at various places along the coasts of the United States; in England; on the shores of the Baltic Sea; in Holland, Belgium, and France. Similarly they are produced on the shores of large lakes or inland seas; the southern end of Lake Michigan is fringed with high sand dunes. They are also of common occurrence along broad river bottom lands in semiarid regions, as along the Arkansas River in western Kansas. Sand dunes are especially abundant in arid regions, where disintegration is rapid and a protective mantle of vegetation is

lacking. Large areas in the great deserts of Africa, central Asia, Arabia, Australia, and western America are covered by them. It is a mistake, however, to suppose that all desert lands are mantled deeply with sand. Even in the Sahara there are great areas of barren rock from which the wind sweeps all loose material.

Shape of Dunes. — The shapes of dunes vary according to the intensity and direction of the wind. Some are linear, and others very irregular; many are curved or crescent-shaped in plan and are called *barchanes*. The windward side of a dune has a gentle slope which is determined by the strength of the wind. The leeward side, capped by a rather sharp crest, has a steeper slope, which is the angle of repose of free falling sand, and varies from about 20° to 30°. The shape thus tells the direction of the prevailing winds.

Migration of Dunes. — A dune once formed is never stationary unless it meets an object larger than itself, or is held by a mat of vegetation.

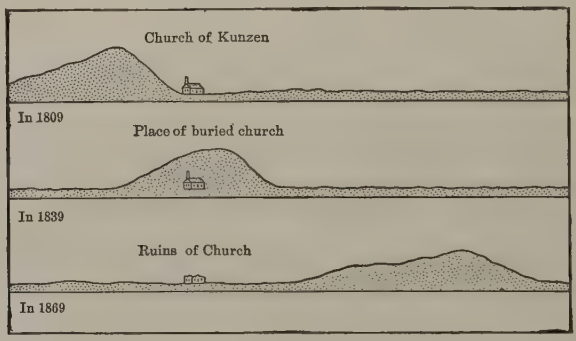

Fig. 20. — Movement of a sand dune during 60 years on the east shore of the Baltic Sea at the village of Kunzen. (After Berendt.)

It moves ceaselessly onward with the prevailing winds, by the slow transference of sand grains from the windward to the leeward side. During this progress its height is maintained or increased, though its outline may suffer change. Thus, along coasts where the prevailing winds are from the sea, dunes may form a belt extending several miles inland. Their rate of movement may average about 20 feet a year, as in Denmark, or reach more than 100 feet a year, as on the Biscayan coast of France. In their march they cover and destroy arable lands, forests, and even towns, leaving desolate sandy wastes behind (Figs. 20, 21). In the deserts of central Asia, Sven Hedin, the explorer, found ruined cities of an ancient civilization emerging from sand dunes which for a long time had overwhelmed them and the fertile lands that once supported them.

The devastating migration of dunes has been mastered along the coasts of European countries by the skillful planting of trees or other vegetation to prevent the removal of the sand particles from the windward slope. The dunes then become stationary. In France the planting of forests on the dunes resulted not only in preventing dune migration but in the raising of profitable forests on otherwise waste land. In arid regions, however, where vegetation will not grow, the migration

Fig. 21. — Forest overwhelmed, killed, and then left exposed by marching sand dunes. Manitou Island, Lake Michigan. (U. S. Geol. Surv.)

of dunes is almost impossible to control, and nothing should be built in the path of their travel.

Loess. — Extensive deposits of fine clay-like materials have been built up chiefly by the wind to form *loess*. The name comes from Alsace, where it designates a peculiar fine-grained, yellowish-brown, lightly compacted earth. It is remarkably fertile. The deposits show no horizontal banding, but form upright bluffs because of a tendency to cleave vertically. These deposits are extensive along the Rhine, occurring well up on the mountain slopes, and form the most fertile soils of central Europe. They cover tens of thousands of square miles in the central Mississippi Valley region, especially in Iowa, Nebraska, and Kansas, and are found also in Oregon and Washington. Similar extensive loess deposits occur in the rich pampas of the Argentine. They

are thick deposits that have formed chiefly from accumulations of dust, which in America and Europe was probably supplied by the flood plains of glacial streams during the recent ice age. There was no vegetation to hold the soil at that time, and the wind blew away the finer particles.

The greatest development of loess is in Asia, particularly north central China, where it covers 230,000 square miles, and the yellow earth washed down from it has given the Yellow River and the Yellow Sea their names. It is believed by Von Richthofen to have been carried by the wind from the great deserts of the interior, especially the Gobi desert, and to have spread over the adjacent area. Its thickness reaches 300 feet or more. It is remarkably fertile, and even though it has been cultivated for thousands of years, it is still productive. Streams have cut canyons into it, and at the base of the bluffs the humbler Chinese have fashioned in it cave dwellings which they have used for centuries. Where roads cross it, the material loosened by the cart wheels has been blown away, and this going on for centuries has caused the roadways to sink lower and lower, so that some of them are now small canyons.

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CHAPTER IV

RAIN AND RUNNING WATER

The Rainfall. — Probably there are no parts of the Earth's surface on which rain does not sometimes fall. Death Valley, in California, one of the driest places in North America, has about 2 inches of rainfall annually. Even the most arid spots in the Libyan Desert have had rain at least once within a measured span of 12 years. But the amount of rainfall a country receives is dependent on a variety of factors, such as the direction of the prevailing winds, the nature of the places over which they have previously passed, and the height above the sea of the country which receives them. Thus it happens that the amount of rainfall received by the land is very unequally distributed over the world; in many places, as in Central America, it may be as much as 100 inches per year and in parts of India 500, but in the great deserts it is less than 10. In North America, it is true in general that in the Atlantic seaboard region and in the southern states the rainfall is 40 inches or more per year; westward to the Mississippi River it diminishes to 30 inches or somewhat more; in the Great Plains region to 20 or less; in the Basin and Range country between the Rocky Mountains and the Sierra Nevada to 10 or less. Locally in the mountains it is increased, because mountains are great condensers of moisture. On the Pacific coast it increases again. Roughly speaking one may term as *arid* those regions where the rainfall is less than 10 inches, as *semiarid* those where it is 10 to 20 inches, and as *humid* those where it is more than 20 inches.

Various things happen to the rainwater after it reaches the surface of the land. A part is trapped by surface depressions and remains in lakes and ponds. A part evaporates, a part sinks below the surface and becomes ground water, and a part flows down the slope of the surface, forming the *immediate run-off*.

The Run-off. — The proportion of the rainfall that contributes to the run-off in a given region depends on several factors such as (1) surface slope, (2) porosity and solubility of surface material, (3) character and amount of vegetation, (4) temperature and humidity of the atmosphere, and (5) distribution of the rainfall throughout the year. It therefore varies greatly from place to place. Certain areas of very soluble and porous limestones have essentially no run-off, all of the rain-

fall sinking below the surface, whereas on steep slopes in the Appalachian Mountains, nearly all of the water from rains in the early spring flows off over the surface, because the ground is already saturated with melted snow. On an average, about one-fifth of all the rain that falls forms run-off. Since much of this fifth flows from areas thousands of feet high to the sea, the power of the resulting streams to do work is very great. As the water under the influence of gravity flows off over the surface during and after rains, it seeks the shortest and steepest route downward, following the lowest depressions wherever they present themselves, and is continuously augmented by seeps and springs representing the reappearance of the ground water which sinks below the surface at higher levels.

Stream Valleys; Stream Erosion. — Every observer, as he stands on the brink of the Grand Canyon of the Colorado and looks down at the winding ribbon of the river a mile below, is impressed with the vast size of the rock-walled valley and with the puny appearance of the stream at its bottom. The earliest observers thought that the Colorado had found this low course already prepared for it by a great gash or rift in the Earth's crust. But the visitor of today, if he knows some of the principles of geology and understands rivers and their behavior, can believe that the river once flowed through a shallow trough at the level of the canyon's brink, and that very gradually the sediment-laden water cut down little by little into the underlying rocks, sawing ever more deeply until in the course of time it dug a trench a mile deep. This same visitor surveys from an eminence the intricate pattern of the canyons and sub-canyons which empty into the main valley, arranged symmetrically like the veins in a leaf, most of them entering flush with the main stream. He realizes that they are not fortuitous, but are all parts of a unified and interrelated system developed in obedience to a common law, and even having a common destiny. And when he sees this same pattern repeated again and again in stream systems large and small throughout the world, he concludes that all rivers act according to certain definite rules imposed on them by their surroundings. The fundamental cause of stream flow is the pull of gravity upon mobile liquid at the Earth's surface. But as water flows it picks up loose rock particles, carries them downstream, and by their aid wears away the solid rock of its bed. This process of wearing away is called *erosion*, whether the active force be a stream, a glacier, the wind, or some other agent.

Development of Streams. — It is probably true that no part of any continent has escaped inundation by the sea at one time or another during the Earth's history. It follows that not infrequently areas of shallow sea bottom (such as the bottom of Hudson Bay) have been

warped up above sea level and have become dry land. As soon as such an area appeared above water, it would have received rainfall, and the run-off would have been shed from the high places toward the low, and so into the sea. All the conditions necessary for stream erosion would be fulfilled — a quantity of water being pulled downward by gravity, with loose mantle rock constantly being formed by weathering processes, ready to be picked up, carried away, and deposited at lower levels by running water.

Consequent Streams. — Consider the history of a land area newly emerged from the sea. During the first rainfall following emergence, the impact of raindrops upon the surface loosens the exposed fine particles of mantle rock and the run-off carries them down the nearest slopes in little temporary rills or rivulets. Where rivulets join at the bases of converging slopes the wash of their combined volumes picks up more loose particles and digs out *gullies*. Adjacent gullies join at lower levels and the constantly increasing volume of water excavates *ravines*. These in turn join to form even larger *valleys*. In this way, with many more junctions and convergences, the run-off reaches the sea in the form of streams loaded with sediment picked up at higher levels. The slopes down which the run-off flows in this early stage are called *initial slopes*, and the streams following or " consequent upon " these slopes are known as *consequent* streams. The rate of flow of the consequent streams depends entirely upon the steepness of the initial slopes which in the present case are gentle because they represent an upraised sea floor. Steep or gentle, however, slopes are always present; and once they have instituted a stream system, the streams will enlarge their valleys and develop new tributaries, actively eroding the land mass until they have brought the entire land surface nearly to the level of the sea.

Since erosion is the wearing away of rock material, erosion by streams therefore involves (1) mechanical wear by rock fragments carried by flowing water; (2) solution of the rocks in stream beds; (3) picking up of worn particles; and (4) transportation of the mechanically worn and dissolved substances. Since weathering is practically universal the comminuting action of the weathering process greatly aids erosion by providing streams with an abundance of loose material for ready transportation. However, a certain amount of deposition of transported material is always going on even in the smallest gullies where erosion is rapid, and thus deposition is a universal companion process to erosion.

The rivers then are the main channels of drainage, and they are the chief factors in carrying away the waste of the land. They are the great trunk lines of transportation for the rock débris delivered to them by their tributaries. In addition they are themselves powerfully capable

of wearing away the land through which they flow; in them the work of running water as a geological agency is most conspicuously displayed. If we should think of a typical river we should imagine it rising in lofty mountains through the union of many impetuous streams or dashing torrents; gathering headway it rolls rapidly through the belt of lower hilly country and emerges upon wide plains through which it wanders in many curves in a quiet and steady flow to the sea. The slope of its bed, its *gradient*, which may be as much as 20 to 30° at the head, becomes less and less until it is nearly horizontal at the river's mouth.

Although we think of this as the ideal course of a river (and it is typical of many of the great rivers of the world such as the Amazon and the Ganges, and of a great number of smaller ones such as the Po in Italy), we constantly find variations from this type. Thus the Mississippi does not rise in a mountainous country, but in a moderately elevated region of low relief, and it has a very low gradient to the sea. On the other hand, some rivers rise in mountains near the sea, and hence their lower plains districts, corresponding to *b* of Fig. 22, may be short or wanting. In North America, the rivers of the northern Atlantic coast are mostly between these extremes and their courses lie between *a* and *b*; those of the southeastern states are more nearly typical, since they rise in the Appalachian Mountains and flow out upon the Atlantic coastal plain.

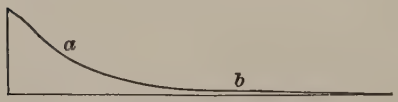


Fig. 22. — Profile of a normal river showing its gradient.

ANALYSIS OF STREAM EROSION

Associated with the destructive or *degradational* process of erosion is a constructive or *aggradational* process of deposition of the eroded material. Thus the normal stream gradient (Fig. 22) is a concave curve partly because erosion is most active near the headwaters of a stream while deposition takes place chiefly in the lower reaches. Strictly speaking, erosion by streams consists of three distinct processes, (1) *corrasion*, (2) *solution*, and (3) *transportation*. The static process of weathering (Chapter III) usually precedes erosion, preparing rock waste for easy seizure by running water, and deposition always follows it. But these two processes are entirely distinct from erosion proper.

1. Corrasion

Stream erosion is limited in its operation, being confined to the bottoms and sides of the channels through which the water passes. A river may be compared to a sinuous, flexible, and endless file, ever moving

forward in one direction, and by means of the moving sand or gravel rasping away the rock beneath and beside it, thus cutting an ever-deepening trench. This particular phase of a river's work is called *corrasion*. The effectiveness with which a river corrades depends on several closely related factors; (1) on the abundance and character of the tools with which the river has to work, (2) on the velocity of its current, and (3) on the nature of the bedrock with which it has to deal.

(*a*) **The River's Tools.** — Clear water moving over rock surfaces has little erosive effect. It has a certain solvent power and may thus

Fig. 23. — River bed full of more or less rounded boulders, showing the tools with which the stream works. Big Creek, Haywood Co., N. C. (U. S. Geol. Surv.)

slowly dissolve and disintegrate rocks, but in order to corrade, a river must have tools. These are supplied by the sand and silt which it carries, and by the gravel and pebbles it can move if swift enough, either in its normal flow, or in times of flood (Fig. 23). This material is supplied to the river chiefly by rain wash and by its tributaries, but the stream also obtains it directly by wearing and undermining the sides of its channel. If its banks are steep, or cliff-like, the débris which accumulates at the foot of such slopes is seized by the river and used as its tools. It is by the striking, bumping, and grinding action of this material,

carried along by the current, that the river is able to cut away the rocks over which it runs and thus to deepen its channel.

In this process the material carried by the river is itself necessarily worn, has its sharp angles removed and becomes rounded or spheroidal, — a form characteristic of the river's tools. Thus, if we find river gravel which consists of hard, well-rounded pebbles, we infer the material has been transported a long distance; on the other hand if the gravel is composed of angular bits of rock, and its situation shows that it has been transported, we infer that the distance must have been short.

Up to a certain point an increase in the amount of grinding material supplied to a river with a given velocity of current aids in its corrasive power. Beyond this point an increase is not effective for the reason that the strength of the current is so consumed in the operation of transporting that the check, given by a tendency to corrade, would cause the river to deposit instead of carrying material farther. Since corrasive power depends on the strength of the blows struck by the moving particles, it is clear that this in turn depends upon the momentum, that is, upon their mass multiplied by the velocity. Hence for a constant velocity, the greater the mass of the particles — that is, the larger and heavier they may be — the more effective agents of erosion they become. Thus in a stream carrying intermingled grains of sand and dust-like particles of clay the sand is the most effective agent.

(*b*) **Velocity and Corrasive Power.** — It is obvious from the preceding paragraph that other things being equal, the swifter a current is the more rapidly it will corrade. For, in a given time, not only will the number of corrading particles passing over a rock surface be increased with a swifter current, but the fact that each particle is moving more rapidly will add to its effectiveness. Further, since a swift stream can carry larger particles than a slower one, and since these larger particles, owing to their momentum, strike the channel with greater force, it appears that a moderate increase in the velocity of a stream will result in a great increase in that stream's cutting power. Calculation of these complex factors shows that doubling a stream's velocity increases its corrasive power at least four times, and perhaps in some cases as much as sixty-four times. In other words, corrasive power varies by a factor between the square and the sixth power of the velocity.

(*c*) **Character of the Bedrock.** — Such rivers as the Platte and the Missouri on the Great Plains, the James, the Roanoke, and the Savannah on the Atlantic Coastal Plain, the Yangtse on the coastal plain of China, the Thames in the London Basin, the Seine in the Paris Basin, and a host of others besides — all are turbid with sediment. They are alike in that they flow through regions of shales, clays, and soft sandstones.

On the other hand most of the streams of New England and the Adirondacks, and of the Highlands of Scotland, together with such streams as the upper Danube, are relatively clear. And these are alike in that they drain regions of hard igneous and metamorphic rocks. Thus we find countless illustrations of the fact that the composition of the rock in a stream valley has a great influence on the rate of erosion. The structure of the rock masses also exercises a profound effect. If they are jointed and thinly bedded, many planes of weakness are present along which the stream can chip out and dislodge fragments. Vertical beds of shale present ideal conditions for this kind of erosion. If the rocks are massive, like granite, streams traversing them erode with difficulty.

2. Solution

Even without visible tools running water wears away the land by solution. Its solvent action is greatly aided by substances already in solution, inherited in part from the ground water that issues upstream in springs and seeps and in part directly from the rainwater which commonly contains dissolved gases acquired from the atmosphere and from decaying vegetation. The rock most susceptible to the solvent action of stream water is limestone (calcium carbonate). But even where a stream flows over relatively insoluble rocks, certain soluble minerals are decomposed, thus loosening the adjacent insoluble mineral grains and preparing them for mechanical seizure by the current.

3. Transportation; the River's Load

The material carried by a stream forms its *load.* While the greater part of this is carried (*a*) mechanically in suspension, a very considerable portion is transported (*b*) chemically in solution, and still another part is (*c*) rolled or moved along the bottom. The ultimate goal of the river is the sea into which, at the end of its journey, its remaining load is deposited. The various aspects of this work demand consideration.

(*a*) **Material in Suspension.** — The size of the particles that a river is able to carry in suspension depends on (1) the character of a river's current, (2) its velocity, and (3) the relative weight or specific gravity of the particles.

1. *Character of the Current.* — If the mass of water forming the current moved forward in a perfectly uniform manner, each particle of water from side to side and from top to bottom moving forward with the same velocity as every other particle, only the very finest material, such as microscopic granules of clay, would remain any length of time in suspension. A sand grain dropped into the stream would sink to the

bottom and there remain at rest, unless the stream were strong enough to roll it along. But the current of streams is not of this character. The more central portions are moving more swiftly, sliding over those toward the bottom and sides, while there is a constant interweaving of swifter sub-currents up and down and toward the sides and even backward, forming eddies or whirling movements. The whole effect is like the stirring of water in a glass. Sand at the bottom is quickly lifted and kept in suspension by these movements while it is carried along by the main current. When particles in suspension in water attain a certain degree of fineness their settling, even when the water is still, becomes very slow. Thus water such as that of the Mississippi may remain turbid for many years.

2. *Velocity and Transportation.* — The velocity of a current depends not only on the gradient, but also on the volume of water involved. Thus of two streams having similar gradients and form of channel, the one having the larger volume of water will have the swifter current. It is also well known that the swifter a current, the larger and heavier the masses it can transport. It has been found that a current running a fifth of a mile in an hour will carry fine clay; one running half a mile in an hour will transport sand; one of a mile an hour will roll along medium-sized gravel, and one of 2 miles an hour will sweep along pebbles the size of an egg. Reduced to mathematical form it may be stated that *the maximum size of particles that a stream can move varies as the sixth power of the velocity.* If the velocity of a stream be doubled it can move particles sixty-four times as large as before. With low velocities of less than a mile an hour and with fine particles this increase does not seem very striking. Thus sand grains may be a hundred, or a thousand fold, as large as those of fine silts or muds and require a doubled or trebled velocity to move them. But with increasing speeds of miles per hour the effect becomes very marked. This explains why rapid streams of 5 miles per hour are able to move small boulders, and sudden floods in narrow valleys, caused by torrential downpours of rain or the bursting of dams, are able to carry with them huge masses of earth and rocks, sweep away bridges and other structures, and cause great damage (Fig. 24). When the St. Francis dam near Los Angeles gave way in 1928 and flooded the valley below, huge blocks of concrete, one of them weighing 10,000 tons, were displaced by the escaping waters.

3. *Effect of Specific Gravity.* — The size of the particle that a stream of a given velocity is able to carry depends also on the specific gravity of the materials composing the particle. A familiar example is that a lead sinker is able to remain at rest on the bottom of a stream which carries away pebbles of an equal size. A practical application of im-

portance is found in the fact that in placer gravels, from which gold is extracted, fine particles of the precious metal are mixed with vastly larger ones of sand and gravel; the water, on account of the high specific gravity of the gold, being able to transport it only with difficulty. Hence if the gold grains are relatively angular, that is, unworn, it is inferred that they cannot have been transported far from the rocks which originally contained them and from which they were derived by erosion. When a prospector finds angular grains of gold, he searches the adjacent slopes for gold-bearing veins. The specific gravity of the

Fig. 24. — In times of flood a stream is able to carry masses and perform work vastly greater than under ordinary conditions, as here shown by the results of flooding. Manti Creek, Utah. (U. S. Geol. Surv.)

great mass of material obtained by erosion and carried by rivers lies between 2.5 and 3.0, that is, it is that much heavier than an equal volume of water. Also the principle that a body immersed in water loses weight equal to that of the water displaced, greatly aids the transporting power of the stream.

(*b*) **Transportation of Material in Solution.** — All river waters carry in solution salts of various kinds that have been leached from the rocks and soils of the country which they drain. Although in a measured

volume of what we call fresh water the amount may seem relatively very small, in the aggregate the weight of material thus dissolved from the land and carried into the sea is enormous. It has been estimated that nearly 2,735,000,000 metric tons[1] of solid substances are thus annually transported into the oceans. The Mississippi carries about 136,000,000 tons, the Connecticut, a small river, 1,000,000, the Danube over 22,000,000, the Nile nearly 17,000,000. In the Mississippi the amount carried in solution is more than a third as large as that carried in mechanical suspension, the quantities being

340,500,000 tons in suspension;
136,400,000 tons in solution.

The most important of the substances thus dissolved and transported are calcium and magnesium carbonates ($CaCO_3$ and $MgCO_3$); calcium, sodium, and potassium sulphates ($CaSO_4$, Na_2SO_4, and K_2SO_4); sodium chloride (NaCl); and silica (SiO_2). In humid regions the abundant vegetation by its decay generates carbonic acid, and by the aid of this the percolating waters dissolve carbonates from the rocks. Hence in humid regions the waters of rivers like the Potomac and the Delaware, have chiefly carbonates in solution; in arid regions where vegetation is sparse or wanting, the waters contain mostly sulphates and chlorides, as in the Colorado and the Rio Grande.

(*c*) **Transportation on the River Bed; Traction.** — In addition to the material carried in suspension and solution a considerable part of the river's load is pushed or rolled along the bottom. What proportion of the whole this may be cannot be accurately determined; it has been thought that in the case of some rivers it is greater than the amount carried in suspension. From studies made on the Mississippi, it is roughly inferred that of the material which it carries into the Gulf of Mexico about 10 per cent or more consists of coarser débris moved along the bottom. It is obvious that, other things being equal, the steeper the gradient a river has the larger will be the amount of the material so moved.

By observation of experiments in troughs it has been found that the amount of material moved by actual sliding or rolling of the particles along the bottom of a stream is much smaller than the amount which progresses by a series of short leaps. Near the bottom of every stream there is a zone filled with grains moving in this fashion (*saltation* = jumping), and above this is the material in suspension. The particles urged forward by sliding, rolling, and saltation are said to be moved by

[1] Metric ton = 1000 kilograms = 2204 pounds.

stream *traction*. The amount carried by traction, compared with suspension, varies with the swiftness of the current and size of the grains.

Manner of Transportation. — In considering the manner in which material is carried one must recall that it is only in swift streams and the upper rapid tributaries of great rivers that boulders and coarse gravel are moved, especially in times of flood. As one goes down a great river the size of the material steadily grows less with diminishing gradient. This is seen not only in the matter in suspension, but on the bars and beaches where it is temporarily deposited. Finally, in those rivers which wander through wide plains before they reach the sea, only the finest sands, silts, and clays are discharged into the ocean, and no coarse material is seen, except chance pebbles and boulders that have been floated downstream in masses of river ice or among the roots of drifting trees.

Nor is the journey a steady or uninterrupted one. The gradient changes from place to place and with it the velocity and transporting power. Material carried down one reach is deposited at the foot of it, while at the head of the next, rapid erosion is cutting the channel headward and material is thus again set in motion. Matter dropped during a season when the current is slack is seized and again hurried forward with the renewed strength that comes in times of flood. Thus, with many waits and pauses, and growing finer by attrition, the mass of material upon which the river works is urged ever forward and onward down stream.

Rate of Erosion; Denudation

It has been estimated that the amount of material in suspension, in solution, and rolled on the bottom, discharged each year into the Gulf of Mexico, if gathered together would form a right-angled prism with a base 1 mile square and a height of 250 feet. If we reckon the whole basin of the Mississippi and its tributaries as covering 1,265,000 square miles, and consider only the material in suspension and solution, it can be calculated from the given data that the entire basin is being lowered at the average rate of 1 foot in 6000 years. An estimate for the whole United States, based on measurements made on its rivers, is about 1 foot in 9000 years. The actual rate is probably greater, because the amount moved by traction is not included; and the two rivers, the Mississippi and the Colorado, which together transport about 80 per cent of the total material taken from the United States each year and delivered in suspension into the sea, are also those which must move the most by traction. Older estimates for the Mississippi basin have been as low as 1 foot in 4000 years, or even less. The rate for its basin

must be faster than that of the United States as a whole, because certain large desert areas contribute very little to the annual run-off.

The gradual lowering of a land surface through a long period of time is referred to as *denudation*. From the foregoing statements it is obvious that we cannot estimate the rate of denudation with any accuracy, but the results are of interest and importance because they indicate the order of magnitude of the figures concerned. We may say, with some confidence, that the area of the United States is being lowered at a rate of 1 foot in from 5000 to 10,000 years, and probably between 7000 and 9000, and where rock many thousands of feet thick has been removed by denudation, we get some notion of the immensely long periods of time involved in the process.

Other rivers, according to circumstances, have given different figures. Thus it has been calculated that the Ganges erodes its basin at the rate of 1 foot in about 1750 years. But its basin culminates against the loftiest mountains in the world and the river has a proportionately rapid descent and erosive power. The basin is also subject during part of the year to a very heavy rainfall and great floods. Consequently the rate is far greater than the average. On the other hand desert regions, like those in central Asia or the Sahara in Africa, with very little rainfall, are eroded with great slowness, the chief agent of transport being the wind. The average height of North America above the sea has been roughly estimated as 2000 feet; at the rate of 1 foot in 7500 years it would take 15,000,000 years to reduce it to sea level; but as erosive processes (excepting solution) go on more and more slowly as the slope is reduced, this time in reality would be enormously lengthened out.

CONSTRUCTIVE WORK OF RIVERS: STREAM DEPOSITION

So far in the study of rivers we have considered the destructive, erosional work they perform — work done chiefly in their upper reaches and seen in the valleys they excavate in the higher lands. Some rivers have a swift course through elevated tracts of country to the sea, their work is cut short when they enter it, and they deposit their load at once; but many, and especially the larger rivers of the world, descend into wide lowlands, through which with steady current they wind to their journey's end. In these lowlands, and at the rivers' mouths, the work done is different from that in the upper reaches; it is largely constructive, rather than destructive, and consists mainly in the deposition and slow shifting of the burden assumed through erosion in the higher part of the course. However, the work of cutting, especially against the valley sides, proceeds along with the work of deposition. It is not possible to separate the two processes in a comprehensive discussion.

Meanders. — As a stream reaches the middle and lower stretches of its valley its velocity constantly diminishes owing to decreasing gradient (Fig. 22). In consequence, it is easily turned by obstacles that it lacks the power to remove. Any deviation from a straight course throws the

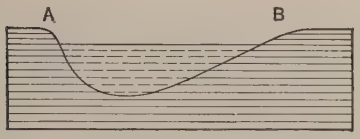


Fig. 25. — Section of a stream channel from *A* to *B* shows it to have the profile seen above, the deepest part lying close in to the bank at *A*. See Fig. 26, *A* and *B*.

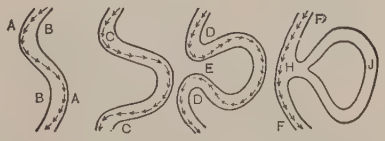


Fig. 26. — The formation of meanders and ox-bows.

current against one of the stream's banks. Some erosion of the bank occurs, and the current is thrown against the opposite bank (as in *A*, Figs. 25 and 26) where again erosion takes place. Symmetrical curves are thus formed, and a continuation of the process increases them. Meanwhile the current is slackened at *B* (Fig. 26), deposition occurs

Fig. 27. — Meanders of a stream in nearly flat region. Trout Creek, Yellowstone Park. (U. S. Geol. Surv.)

there, and the inside of the curve is built out as the outside is cut away. In this manner the arcs of the curves become more and more pronounced as *CC* and *DD*. Eventually a loop, as at *E*, is cut through, leaving an island in the stream, the main current takes the shortest route *FF*, and the entrances to the abandoned channel are quickly silted up, leaving a

shallow crescentic lake. The symmetrical curves are called *meanders* (from the River Meander in Asia Minor) and the resulting lakes, *oxbow lakes*. Meanders are roughly proportional in size to the size of the stream with which they are associated. The arcs in small streams commonly have circumferences of only a few score feet, but those of the Mississippi may be 20 miles or more in length.

Lateral cutting is necessarily more effective against those portions of the meander curves which face upstream than against those which face downstream; hence each meander loop tends to shift slowly downstream during its existence. Thus a constant succession of meanders moves almost imperceptibly mouthward past any given point in a stream valley. This steady shift downstream is called *sweep*.

Alluvial Flats; Lateral Planation. — As a meandering stream wanders from side to side in its valley, it impinges against the valley bluffs from time to time (*A*, Fig. 28), cuts them down by undermining, and carries the material away. By continuation of this process the originally narrow valley is widened. The work is known as *lateral planation*. Furthermore, by continuous deposition on the inside bank of every meander curve, the stream eventually covers the entire floor of its valley with material dropped from its load. Whether the component material (river alluvium) is coarse gravel or fine silt, the resulting narrow plain of deposition is called an *alluvial flat*.

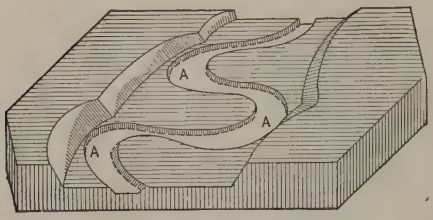


Fig. 28. — A river widening its valley by meandering and planation.

Flood Plains; Natural Levees. — As the gradient of a stream steadily lessens, and as lateral planation increases, the sediment dropped is spread continuously along the valley bottom, building up great flats. In the spring, melting snows and heavy rains pour a vast volume of water into the tributaries of the stream. This volume, concentrated in the main stream, raises the water level until it overtops the channel and floods the bordering flats. The water which has been moving swiftly through a deep channel, with friction at a minimum, thus suddenly forced out on to a shallow flat, is quickly and almost completely checked. Concentrated deposition therefore takes place along the immediate borders of the flooded channel, grading outward away from the stream into much thinner deposits. In consequence, low ridges are built up paralleling and confining the stream channel; and these remain after the flood has subsided and the stream has shrunk to its old dimensions. They are known as *natural levees* (Fig. 29). Beyond

them the land is low and usually swampy. The whole broad flat, including both the natural levees and the flanking lower lands, is called the *flood plain.* The flood plain of the Mississippi covers an area of about 30,000 square miles, and a large portion of it consists of extensive swamps.

With the subsidence of the flood, the decrease in volume of water flowing through the channel not only causes the stream to cease to corrade (i.e., scour out its channel); it may cause it actually to deposit. Thus during low water the stream silts up its channel with sediment. In this way the stream may gradually come to flow at a level higher than that of the flats beyond the levees, being restrained *only by the levees* from deserting its course and occupying the flats. This is the situation indicated in Fig. 29 and is moreover the condition in the lower Mississippi when in normal flood. This has become a serious matter from the

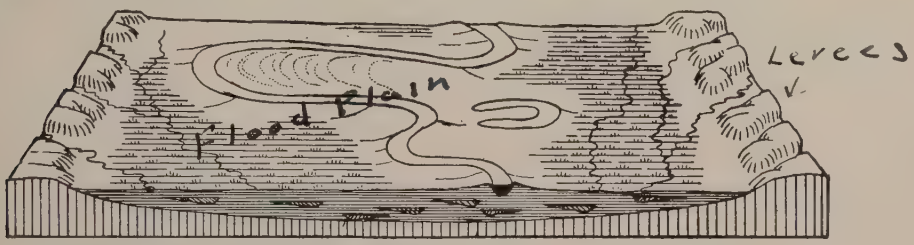


Fig. 29. — Flood plain with natural levees built by a large stream.

point of view of floods, with the gradual settlement and agricultural development of the low, ill-drained areas back of the levees.

Artificial Levees; Floods. — Before the settlement of the lower Mississippi Valley, the river was in the habit of overflowing its banks periodically, the floodwater reaching the sea through the low flats beyond the natural levees. As early as 100 years ago attempts were made to confine the flooded river to its channel, in order to reclaim for agricultural purposes the adjacent rich and fertile alluvial lowlands. These attempts took the form of artificial levees which were built upon the natural ones, thus deepening and heightening the channel at the same time. If a given amount of water is confined to a narrowed course and thus prevented from normal spreading, its channel must necessarily be deepened. A part of this added depth is attained locally by increased erosion of the bottom, but much of it is attained by rise of the water surface. Hence whenever a new levee is put in, all those already in existence must be raised. The first levee, built at New Orleans, was 4 feet high. Today the average height is more than 13 feet. In 1902 there were 1300 miles of levees along the Mississippi; in 1927 there were

2500. The river level is gradually rising higher above the adjacent basins, and when floods do occur, they are correspondingly more destructive, as witness the disastrous flood of 1927. The river has been excluded from the flood plain, a part of its natural domain, and when it overflows its artificially restricted channel it is merely seeking its age-long rights. The building of levees is therefore not in itself an adequate method of preventing floods. The time is approaching when some additional means of combating floods will have to be adopted.

The scheme of building levees has been followed in many parts of the world. In the flood plain of the Po the river bed is above the housetops. Breaks in the levees on the Hoangho at various times have resulted in enormous loss of life. In the flood of 1887 the Hoangho, called "China's Sorrow," drowned considerably more than a million people.

Deltas. — When the current at the mouth of a stream is checked by a body of standing water such as a lake or the ocean, its load is promptly

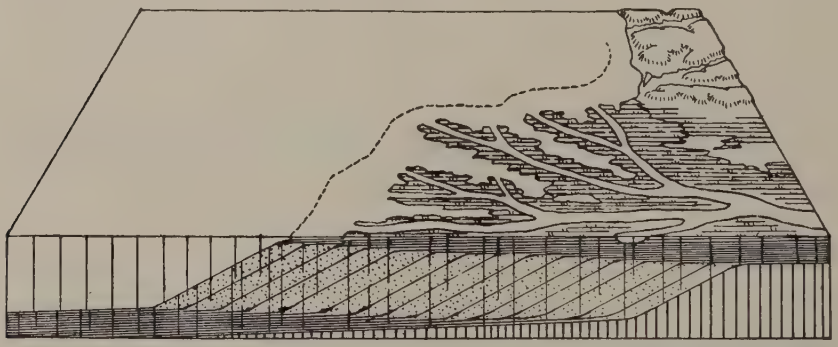


Fig. 30. — Ideal plan and section of a delta. Plan shows delta fingers and distributaries. Section shows thick, steeply dipping foreset beds overlain and underlain by thinner topset and bottomset beds, respectively. Relation of topset to foreset beds indicates sinking of the land as the delta grew.

dropped on the bottom, building an embankment with a front which grows outward like a railroad or highway fill in process of construction across a valley. As the deposit is built up close to the water surface, the flood plain usually encroaches upon it from upstream and gradually covers it so that it is built above water in a crudely triangular shape with one apex pointing upstream. From this shape, resembling the Greek letter, the deposit derives its name of *delta*. This name is applied regardless of whether the embankment remains submerged or whether the encroachment of the flood plain has raised it above the water surface.

The bulk of the delta-forming material is dropped on the frontal slope of the growing embankment, forming thick *foreset beds* (Fig. 30).

Some of the finest sediment however remains longer in suspension and is carried farther out and dropped as fine *bottomset beds* which thin outward away from the delta. Along the top, where erosion and deposition alternate as the stream current changes seasonally, thin horizontal *topset beds* are laid down. The thickness of the whole mass is in some cases very great. Borings put down in Venice, which is built on the delta of the Po, reached a depth of more than 500 feet without attaining the bottom of the delta beds.

Whenever the river in flood overtops the natural levees, it has opportunities to break through at certain points and to flow seaward through some new channel, leaving a diminished volume of water to escape through the old channel. Since this happens so frequently as to be the rule in the case of large rivers left to themselves, new outlets are broken through soon after they have been formed, and a branching system of distributaries grows up, giving shape to the delta. Thus a long stream ending in a delta is like a rope frayed at both ends, the strands at the upper end being represented by the tributaries and the shorter ones at the lower end by the distributaries. The branching system extending seaward is the skeleton of the growing delta; between the long arms lie shallow basins which gradually fill with sediment during floods and thus become low land.

The shifting of the main channel through the development of new distributaries is strikingly illustrated by the case of the Hoangho, which for approximately 700 years prior to 1852 had discharged eastward into the Yellow Sea. In 1852 it broke its banks at a point more than 300 miles above its mouth, formed a new channel northeastward across the great alluvial flats of the province of Shantung, and finally emptied into the gulf of Chihli, almost 300 miles north of its old mouth. The Hoangho has occupied this new course with minor distributaries since 1852, and the old channel has largely dried up.

The Mississippi attempted a similar change in April, 1890, breaking its banks at the Nita Crevasse[1] between New Orleans and Baton Rouge, at a point well over a hundred miles from its mouth. From here it flowed eastward through Lake Maurepas, Lake Pontchartrain, and Lake Borgne into Mississippi Sound, inundating a wide area, causing great damage, and halting railroad traffic for two months. The river at length resumed its old course, after having taught local engineers that in the vicinity of the Nita Crevasse the Mississippi was normally about 21 feet above the level of the bordering swamps and flats.

[1] The term *crevasse* in the region of the lower Mississippi refers to a break in a levee.

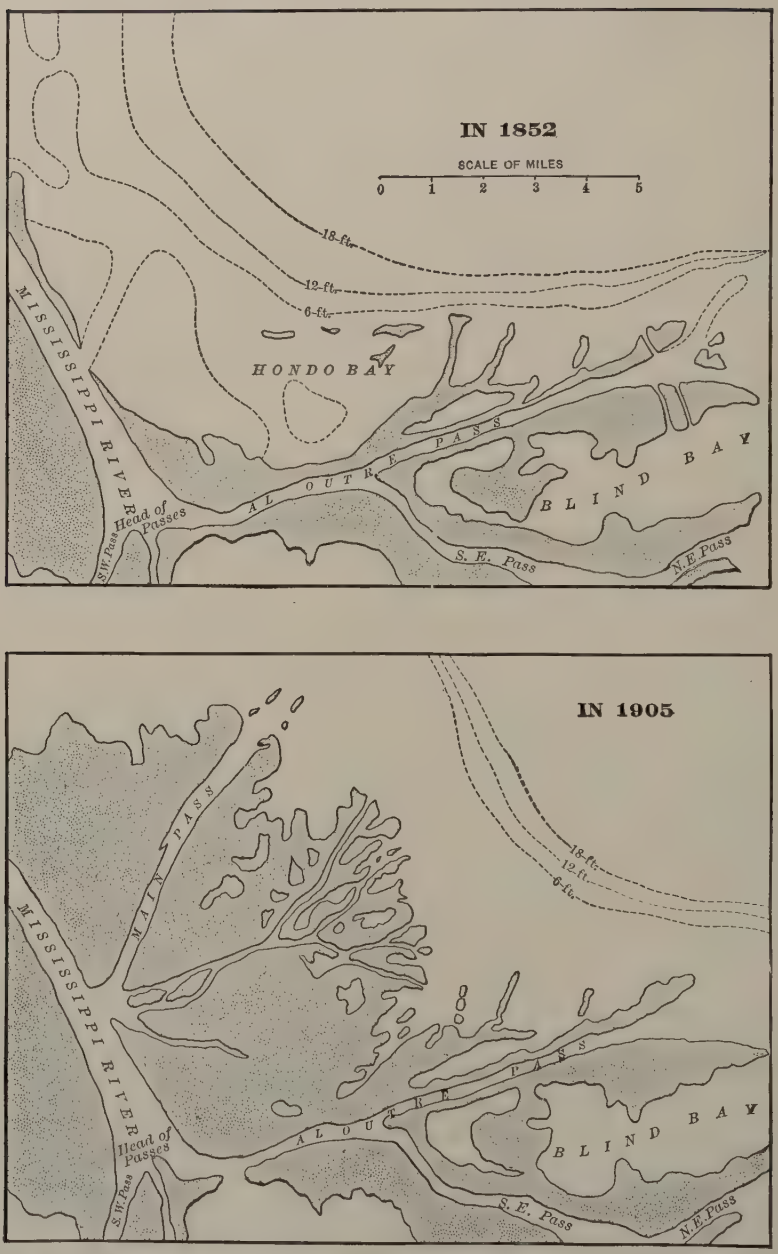


Fig. 31. — Growth of the Mississippi delta during 50 years. (After G. R. Putnam.)

Rate of Delta Building. — In this way the land at the mouths of large rivers is constantly being added to at the expense of the sea. The rate at which this advance of the land takes place is variable, depending upon such factors as (1) depth of water offshore, (2) volume of stream-carried sediment, and (3) power of waves and shore currents to sweep away the newly deposited material. It is estimated that the delta of the Mississippi is pushing forward into the Gulf at the rate of more than 250 feet each year (Fig. 31), and the Po pushes into the Adriatic at nearly as great a rate. Since 400 B.C., the Rhone has been encroaching on the Mediterranean at the rate of about 36 feet annually, whereas the deltas of the Danube and the Nile are each growing only about 13 feet yearly. The obstructions to navigation caused by deposition at the mouths of the Mississippi have been successfully removed by the build-

Fig. 32. — Delta of the Nile, showing its form and distributaries.

ing of extensions of the natural levees out into shallow water. These extensions, called *jetties*, confine the current, so that with increased scour it deepens its channel and at the same time carries its load of sediment out into deep water before dropping it.

Size and Form of Deltas. — It follows from their rapid growth that the deltas of great rivers form large areas of land. The Nile delta (Fig. 32) is nearly 100 miles long and 200 miles broad on its seaward front. The combined delta formed by the Ganges and Brahmaputra is 200 miles

long and has an area of possibly 40,000 square miles. The Mississippi delta is likewise 200 miles long, but is much narrower, having an area of not much more than 12,000 square miles. The Po delta, already large at the beginning of the Christian Era, has increased by nearly 100 square miles since that time.

No two large deltas have the same appearance because of the capricious changes in the distributaries of the rivers and because waves and shore currents erode the deltas at varying rates. The Mississippi delta with its long projections (Fig. 31) built out successively by changing distributaries, shows that deposition by the stream is dominant over erosion by waves and shore currents in the Gulf of Mexico. It is a typical example of the " lobate " type. The Tiber (Fig. 33) has a " cuspate " delta in which shore erosion seems to have the upper hand over stream deposition, while the Nile delta (Fig. 32) is intermediate between the other two and approximates an " arcuate " type (Johnson).

Fig. 33. — Cuspate delta of the Tiber. (Johnson.)

Stream Terraces. — Many stream valleys, particularly in their middle reaches, are bordered by bench-like flats the tops of which are higher than even the flood stages of the present streams. In many cases there are several of them in series, rising away from the river like two long flights of steps facing each other. In some valleys they are continuous for long distances. Such flats, high and dry, are called *stream terraces*. Upon examination of these forms it soon appears that some are made of rock (*rock terraces*) whereas others apparently contain no bedrock but consist entirely of unconsolidated sands, gravels, and clays deposited by the streams when they were flowing at higher levels (*flood-plain terraces*).

Rock Terraces. — These are of two kinds. (1) Rock terraces found usually in dry regions where valleys have been cut down through alternating layers of weak and resistant rocks. The resistant layers, less easily eroded as the stream cuts down, are left standing out in relief, while the weak layers are etched back (Fig. 34). The terraces in the

Grand Canyon are an excellent example. (2) Rock terraces caused by successive uplifts of the land. These are mentioned here only for the

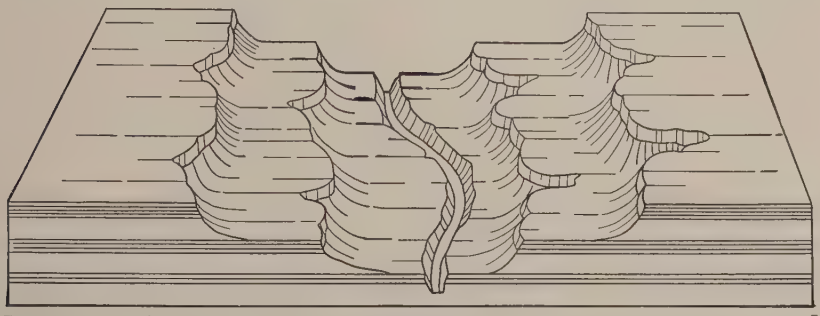


Fig. 34. — Rock terraces caused by differential erosion of a series of strata of unequal resistance.

sake of completeness and are discussed in connection with changes of level (Chapter XVII).

Flood-plain Terraces. — Flood-plain terraces, the most common type, are formed by a stream which has begun to degrade following a

Fig. 35. — Terraces on the middle Fraser River, B. C. (F. F. Osborne.)

long period of aggradation in which a flood plain was built up. It is obvious that any increase in the gradient or volume of an aggrading stream or decrease in its load would cause it to begin to degrade and hence to leave terraces (Figs. 35 and 36). A sudden upward warping of the

land while streams are flowing over it, increasing their gradient, is regarded as a common cause. Another cause not rare in glaciated regions is the melting away of glaciers and the consequent great diminution of the load of the streams that drained them. Many terraces also

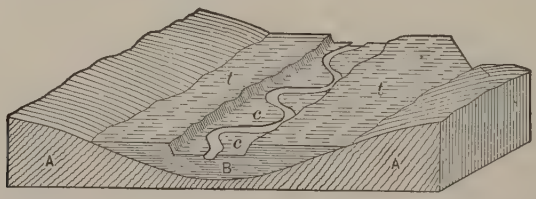


Fig. 36. — Formation of flood-plain terraces. *AA*, section of river-cut valley; *B*, alluvial deposits of river; *tt*, former flood plain, now forming terraces; *c*, new flood plain.

are *rock-defended*. They are developed in this way: A stream of low gradient which has been slowly excavating the deposits in the bottom of its valley, and is meandering from side to side across its valley floor, encounters at numerous points the bedrock of the valley wall. Deflected by this unexpected obstacle, it swings away and is thus prevented

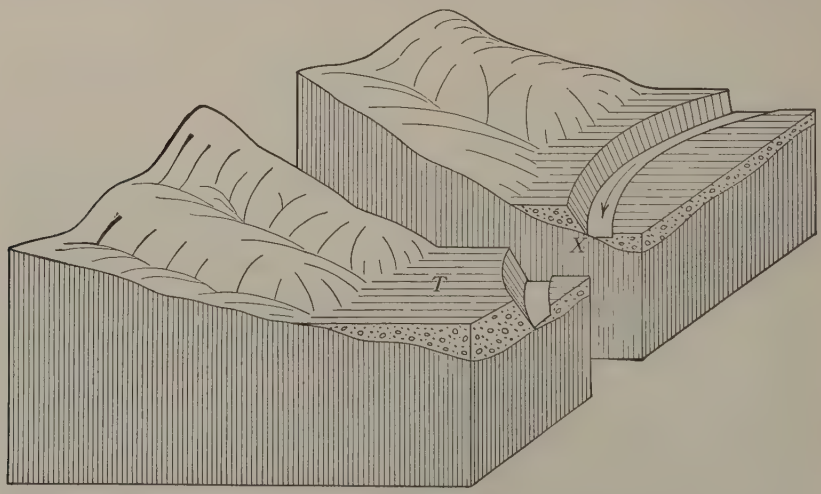


Fig. 37. — Old flood plain "defended" against undercutting by a rock outcrop at *X*, thus gradually forming the terrace *T*. Base of block (vertical ruling) is solid rock.

from undercutting the remnant of the old flood plain above the protective rock outcrop, which remains as a terrace, gradually increasing in height as the stream continues to cut downward (Fig. 37).

Alluvial Cones and Fans. — When a swift tributary stream enters a wide and nearly level valley, the abrupt change in its gradient may

cause it to deposit the greater part of its load on the valley floor. In this way a low semi-conical elevation is formed, radiating out from the mouth of the tributary (Fig. 38). Such forms are generally known as *alluvial fans*, but if steep they are sometimes called *alluvial cones*. In one sense they may be regarded as deltas formed on land, but they differ from deltas in having a sloping rather than a flat top, since their upbuilding is not controlled by the surface of a body of standing water.

Fig. 38. — Alluvial cone, made by a tributary to a larger stream. Stoughton, Wis. (U. S. Geol. Surv.)

An alluvial fan acquires its shape because of uniform distribution of débris over its surface by the parent stream. The stream repeatedly silts up its channel, overflows, and forms new distributaries. Thus when one part of the fan is built up, the stream shifts to another course (which of necessity is temporarily lower) and builds that up. In this way the entire surface is covered by the stream.

Alluvial fans are not common in humid regions. Where present at all they are small, and are usually developed in regions of soft material easy to erode, such for example as glacial deposits. In many dry regions, on the other hand, deposits of this type are characteristically large, and form conspicuous features of arid landscapes.

Structure of River Deposits; Stratification. — All deposits by rivers are so laid down through the sorting activity of water, that they consist of distinct layers or beds of varying thickness. Usually these beds are

very regularly parallel for rather limited distances. Deposits which exhibit this laminated, banded, or bedded appearance are said to be *stratified*, and the arrangement is called *stratification*. In contrast to strata deposited on the sea floor, stream-laid beds have many local irregularities, and exhibit other peculiarities that reflect conditions in a stream valley. Many exposed sections of old rock strata show stratification characteristic of rivers, and we are thus enabled to reconstruct the courses of former streams so old that their deposits have had time to be buried and slowly converted into solid rock.

STREAM VALLEYS

Development of Valleys. — Any land surface on which rain falls sheds the run-off down the lowest routes to the sea. The run-off loosens the surface soil along these low channels, carries away the débris, and thus excavates gullies. And stream valleys are merely gullies grown big. The evolution of gullies into valleys takes place in this way:

Gullies form where the run-off is concentrated. Concentration of run-off enlarges a gully by erosion on its side slopes and at its head. The larger the gully becomes, the more thoroughly it concentrates the run-off, and an endless chain of cause and effect is thereby set up. Erosion at its head causes the gully to lengthen headward, and the slope wash down its sides widens it after each successive rain. At the same time the even more concentrated run-off through the bottom deepens the gully and lengthens it mouthward as well. Thus constantly undergoing enlargement, the gully becomes first a ravine and then a valley. The normal cross section of a valley which is undergoing rapid erosion is that of a **V** (Fig. 39), because the river, occupying a relatively small space, is cutting downward, while at the same time rain wash and gullying tend to broaden the trench the river makes, by washing down the material composing the valley walls. As already shown, as fast as this débris reaches the river, it is seized and carried away. The cross section of a valley depends then on the relative balance between two agencies, downcutting by the river and broadening by weathering and rain wash. Thus in a region where the gradient is steep, downcutting by the river proceeds much more rapidly than weathering and rain wash, and the valley will be deeply incised and have profiles approaching *ara* (Fig. 39). As time goes on and the river gradient is lessened the cutting by the river becomes slower and slower;

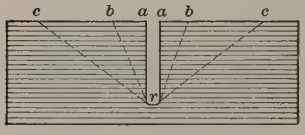


Fig. 39. — Section of a river valley; *ara*, material removed by corrasion; *arb*, *brc*, material removed by weathering and rain wash; river *r*, trenching downward.

weathering then becomes relatively more and more pronounced and the valley widens out as shown in *brb* (Fig. 39).

Intermittent Streams and Permanent Streams. — While in the gully stage, the valley is likely to carry water only during and after rains.

Fig. 40. — A valley in a youthful stage of its history. Yellowstone River. (U. S. Geol. Surv.)

Under these conditions the stream is said to be *intermittent*. As erosion deepens the gully, however, more and more of the rain water that enters the ground directly has opportunity to emerge again in the gully sides, and to contribute to the flow of water on the gully bottom long after the parent rains have ceased. Eventually the gully (perhaps a valley now) is excavated to and below the level at which all openings in the rock are

permanently filled with ground water (Chapter VI). From this time on the water seeps steadily and uninterruptedly into the valley bottom, and the resulting stream is said to be *permanent.*

Baselevel. — When a stream reaches the sea its velocity is checked, its load is deposited, it can cut downward no further, and consequently it can erode no more. The sea level therefore is a level below which streams cannot cut.[1] The level of the sea, projected inland as an imaginary plane below the surface of the land, is called *baselevel* because it is the ultimate base and goal of denudation by streams. It follows that a stream hastens its own downfall every time it removes a cubic yard of material from any part of its bed, because thus by just so much it lowers the gradient on which its velocity depends. Since corrasion and transportation depend on velocity, which in turn depends largely on slope, it is evident that the gradient of a stream is steepest in the early stages of its history and that it progressively decreases as baselevel is approached. In Fig. 22 the base line represents the baselevel toward which the gradient *ab* is steadily being lowered, but which it can never reach.

Grade. — Since in those places where the gradient is lessened a stream tends to deposit, while erosion again sets in when the gradient increases, it follows that, as time goes on, a river proceeds to fill up the hollows and to cut away the projections in its bed and thus to establish a definite gradient. The gradient which the river seeks to establish is one at which, in each part of its course, the velocity is sufficient for the volume of water there present to transport its burden without erosion or deposition; the stream is then said to be *at grade.* This does not mean that the gradient is necessarily uniform from source to sea; it may be relatively much steeper in the upper course, where the load consists of coarse débris and the volume of water is small, than in the lower part where the slope is gentle but the volume of water is large and the load consists of fine sediment. A heavily loaded stream, like the Platte (Fig. 41), may become graded on a relatively steep slope, as compared with one, not fully loaded, which on such a slope would be ungraded and still degrading. The lower parts of great rivers such as the Mississippi become graded while, in their headwaters, cutting and deepening by erosion are still actively going on.

Thus in summary we may say that a stream is *at grade* when its transporting power and the load given it to carry are about equal. It is *aggrading* when the load it has to carry exceeds its ability to transport. It is *degrading* when its ability to do work is in excess of the ma-

[1] The fact that certain rivers can and do erode their lower channels well below sea level does not invalidate the application of the baselevel principle to broad areas.

Fig. 41. — A heavily loaded river. Note the wide bed with many shallow interlacing channels and very broad valley. North Platte River above Gering, Neb. (U. S. Geol. Surv.)

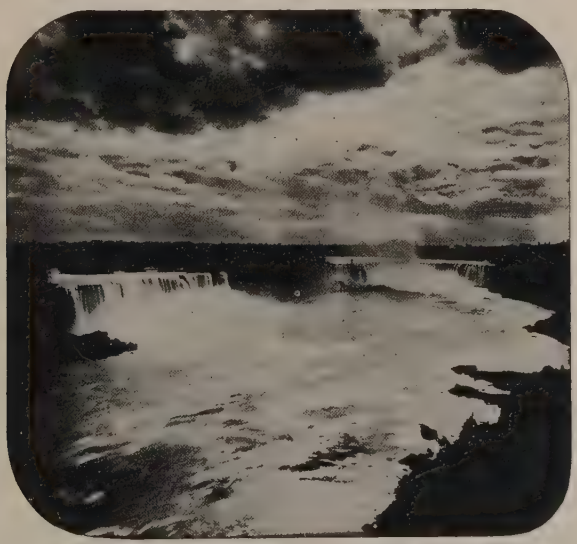


Fig. 42. — General view of Niagara Falls.

terial to be carried, and the excess of energy is employed in deepening its channel. When the stream has reached grade it is said to have reached its "profile of equilibrium" — the profile which permits a transporting power closely adjusted to the amount of waste to be carried.

Relation of Tributaries to Main Streams. — Examination of drainage systems shows that in a vast majority of cases the tributaries of a river enter it at grade, i.e., at the same elevation as the main stream. They are thus said to be *accordant.* The reason for this is that as the main stream cuts down its bed, the resulting increased gradient which is given the tributaries enables them to keep pace in downcutting in spite of the smaller volume of water they contain. But this may increase the ratio of the trenching of the lateral valleys over their widening to a greater degree than in the main trunk valley and hence they grow proportionately narrower and steeper. Examples are afforded by some of the tributaries of Colorado River. In their effort to keep accordant relations with the main stream they have cut narrow slot-like canyons. In some cases, however, in the younger stages of normal valleys, small tributary streams, unable to keep up with a rapidly downcutting river, are obliged to cascade down the main valley walls.

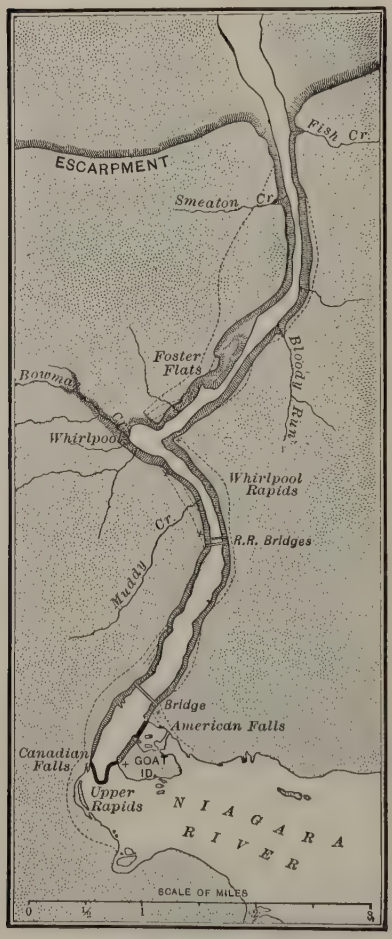


Fig. 43. — Map of Niagara River and Falls. (After G. K. Gilbert.)

Falls and Rapids. — Falls and rapids are common in the valleys of swift streams. The majority of them are the result of the unequal erosion of rock masses composed of both hard and soft layers. This is magnificently illustrated in the great cataract at Niagara. Niagara River, which drains the four upper Great Lakes, in its course of 36 miles from Lake Erie flows over a plateau which terminates near Lake Ontario in an escarpment more than 300 feet high. The plateau is capped by a

resistant layer of limestone under which are soft, easily eroded shales. Originally the falls was situated at Lewiston near the mouth of the river, and falling over the escarpment had its full height at this point. These relations are shown in Fig. 43. By the gradual disintegration and undermining of the softer underlying shale the harder limestone on top is left projecting as a lip over which the water falls (Fig. 44). From time to time this projecting rock, left unsupported and penetrated by joint cracks, also falls and is carried away. By this means the falls maintains itself and at the same time steadily moves upstream, leaving a deep gorge behind it. It is now 7 miles above its original position.

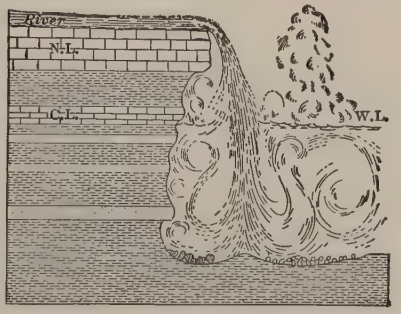


Fig. 44. — Section showing rock layers and cause of falls at Niagara. (After G. K. Gilbert.) N. L., Niagara limestone with soft shale below; C. L., Clinton limestone with shales and sandstones below. 1 inch = 300 ft. W. L. = water level of pool.

The recession of Niagara Falls, and the rate at which it takes place, is a matter of interest and has been the subject of much study because it gives an idea of the length of time involved in geologic processes. Successive surveys made throughout a period of 50 years have shown that the falls is retreating at a rate that averages 5 feet per year. If this rate was maintained from the time the falls first began to be cut, the length of time involved in the cutting of the gorge below the falls (7 miles) would be 7000 years. This is a minimum estimate, but the problem is not so simple as this, since many factors, involving various changes in the river and in the volume of its water during the past must be taken into account. Some estimates which have considered these factors run as high as 35,000 years. Although we do not know the length of time with even an approach to accuracy these estimates are of value in that they show it is to be reckoned in tens of thousands of years, not in hundreds, nor in millions.

Many other famous falls are due to an arrangement of rocks similar to that at Niagara, such as the Falls of St. Anthony on the Mississippi at Minneapolis, and its tributary streams, which fall into the gorge below; Shoshone Falls on the Snake River in Idaho; those on the tributaries of the Columbia River, and many others.

Falls are caused in other ways as well; by glaciers, by the accidental damming of streams back of lava flows and landslides, and by uplift of the land relative to the sea. But whatever their cause, falls cannot indefinitely persist. Increased velocity at their crests results in in-

creased erosion; thus the falls are worn down more rapidly than the reaches above and below them; they pass into rapids and disappear as the streams reduce their valleys to grade.

Potholes. — Circular excavations worn in bedrock by whirling eddies are common in the beds of streams below falls and rapids. They are called *potholes* (Fig. 45). If the conformation of the stream bed is such that an eddy persists in one place, the water whirls sand and gravel with

Fig. 45. — Potholes in granite. Tuolumne River, Cal. (U. S. Geol. Surv.)

it, and this bores downward; although the material wears out in grinding, it is continually replaced by fresh débris, and so the process continues. Potholes have diameters ranging from a few inches up to 50 feet; their depth may be even greater. They are of interest in that they indicate clearly the action of whirling water and, occurring not uncommonly in rock now far from any stream, they prove that at one time this rock was the bed of a rapid current.

THE CYCLE OF STREAM EROSION

If the foregoing conclusions on stream behavior are true; if streams steadily enlarge their valleys, and if tributaries develop and enlarge their valleys, and if cutting by the whole system is limited downward by a *baselevel*, what will be the final result of long-continued erosion by streams on a given land mass? Obviously there can be but one answer: the

wasting away of the land to a low, gently sloping surface from which the water must drain sluggishly to the sea. This process, the complete denudation of a land mass, of course involves an enormous amount of time, but the final result is inevitable. The series of changes involved in the complete reduction of a region to baselevel constitutes a *cycle of erosion.* The time required necessarily varies with varying circumstances such as initial elevation above the sea, resistance to erosion of the underlying rock, and amount of rainfall and run-off.

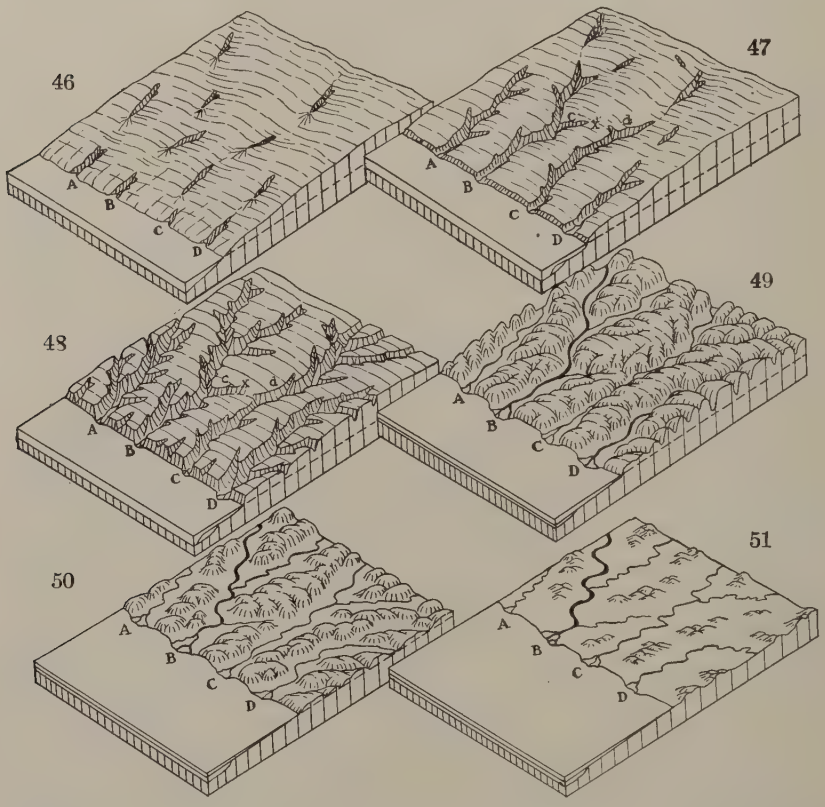
The rainfall factor exercises the chief control over the process of denudation by streams, by controlling both the rate of erosion and the places where at a given time erosion and deposition occur. Thus in regions where rainfall is very slight, the streams dry up before they reach the sea, and all their débris is deposited inland. The stages of the erosion cycle under various types of climate must next be considered.

The Cycle of Stream Erosion in a Humid Region

Although the process governed by the cycle is continuous, it is divided for convenience into stages, as follows: *initial stage, youth, maturity, old age.* It must be remembered that each of these stages grades into the next, and that all are parts of one unbroken chain of events. In connection with the following account of a typical case, reference should be constantly made to Figs. 46–51.

Initial Stage; Consequent Streams. — An upwarped area of sea bottom, bearing on its surface initial irregularities, appears above the sea. For the sake of simplicity let the material composing the mass be broadly homogeneous, with only local variations in its resistance to erosion. Gullies develop (Fig. 46) and grow mouthward and headward under the control of gravity and the initial slopes. Streams whose development is thus controlled by original surface irregularities are called *consequent streams.* Adjacent gullies grow into one another on favorable slopes, forming connected chains. They gradually become ravines and then valleys. In the early stages of the cycle, the gradient is so steep that downcutting of the valley bottoms is dominant over slope wash on the valley sides; hence the valleys and gullies are steep-sided and sharply V-shaped. Any irregularity of slope or material in the side of a gully is apt to concentrate run-off and thus to develop a tributary. Since no gully is uniform in these respects, tributaries rapidly develop, all lengthening themselves headward from the parent gully. When the initial gullies have developed initial tributaries, they break up the continuity of the initial slopes, and the land area passes from the *initial stage* into *youth.*

Youthful Stage. — The land area is now drained by an integrated drainage system consisting of main streams developed from the preëxisting initial slopes. These slopes were so arranged that more concen-



Figs. 46–51. — Ideal cycle of stream erosion under a humid climate and in homogeneous rocks.

Fig. 46. — Initial stage, showing gullies developing wherever the run-off is concentrated.
Fig. 47. — Early youth, showing integration of main drainage lines and growth of the stronger at the expense of the weaker streams.
Fig. 48. — Later youth, showing the reduction of the initial surface to irregular flat-topped ridges.
Fig. 49. — Early maturity, showing dissection of divides into flowing slopes and development of alluvial flats.
Fig. 50. — Later maturity, showing decrease in relief, lowering of slopes and widening of valleys.
Fig. 51. — Old age, showing development of peneplain with monadnocks.

tration of drainage took place along the lines of the infant streams *B* and *C* (Fig. 47) than along streams *A* and *D*. In other words, the sum total of depressions in the areas of *B* and *C* made those areas lower than those of *A* and *D*. The chain of cause and effect was thus set up most

rapidly and thoroughly along B and C and these streams sent out more tributaries and grew headward more swiftly than did A and D. Briefly stated, a struggle for existence takes place among adjacent streams in the competition to grow big and therefore to absorb larger drainage areas. In this case, the streams (B and C) favored by the initial slopes win out and maintain their lead over their less favored and therefore weaker neighbors (A and D).

Divides. — The area of higher land between two valleys is called a *divide.* The divide between two parallel gullies of unequal size may be completely destroyed by the growth of the larger gully (Fig. 52). The

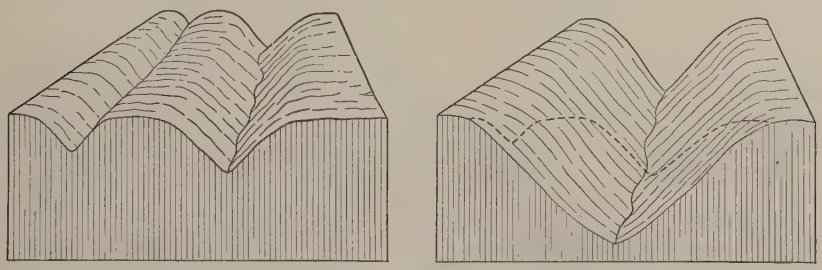


Fig. 52. — Lateral seizure of a small gully by a larger one.

divide between the heads of two streams flowing in opposite directions is easily shifted laterally. Such a situation is developing at x (Fig. 47). Two streams, c and d, tributary respectively to B and to C, have worked headward toward each other, thus narrowing the broad divide which

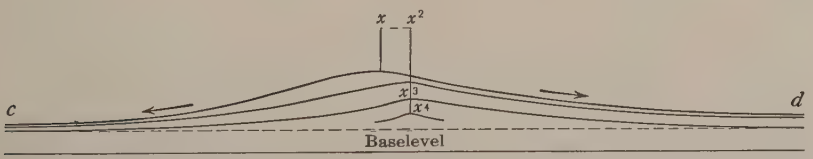


Fig. 53. — Evolution of a shifting divide into a permanent divide. Greater erosion by stream c than by stream d results in lateral shifting of the divide through the distance xx^2 as well as in lowering it. At this stage, erosion has become equal on both sides of the divide so that further downcutting results merely in lowering the (now permanent) divide to x^3, x^4, etc.

formerly separated them. Stream c has a shorter journey to the sea than has stream d; hence the gradient of c is steeper; hence its power to corrade and to transport is greater; hence it eats both downward and headward more rapidly than does d. This inequality in rate of erosion results not only in lowering the divide x but in shifting it away from c toward d (Fig. 53). When erosion by c is equaled by erosion by d the

divide ceases to move laterally and further erosion can only lower it. The divide is then said to be *permanent.*

The streams in the youthful stage have gradients as steep as the height of the land above sea level permits; and since they are energetically cutting into their valley bottoms, the valleys are V-shaped in cross section, with steep sides whose slope depends on the resistance of the rock composing them (Fig. 48). The valley courses are crooked with irregular bends, all of them determined by the initial irregularities of the land surface. Tributaries develop rapidly, their valleys working headward from the main streams like branches growing from the trunks of trees. In fact, so closely does the pattern of a stream system under these conditions resemble a tree with its branches and twigs, that it is called a *dendritic* pattern. Each tributary enters its main valley at a level with the main stream. Downcutting by the tributaries keeps pace with downcutting by the main streams, and thus the whole system is delicately balanced and adjusted throughout its extent. The crooks and bends given each stream in its initial stage deflect the currents from side to side of their valleys, but downcutting is so rapid that no stream remains at one level long enough to allow appreciable widening of its valley by lateral cutting. As the countless tributaries continue to dissect the initial surface, the broad initial divides contract into narrow and irregular ridges. The time of youth is the time of scenic grandeur in a landscape. Deep gash-like valleys (Fig. 40) and canyons with foaming rapids, precipitous cliffs, and high ridges are characteristic. But the very force that sculptured these forms will inevitably destroy them.

It is during the period of youth that landslides are common in regions where steep slopes and suitable climatic conditions are found. The sliding of great masses of rock material from the steep sides of mountains and canyons hastens materially the destruction of the land.

Mature Stage. — When the tributaries have worked headward so far that the narrow ridge-like divides have been dissected into short hills and spurs, and when the valley sides and the tributary heads have destroyed all of the initial surface by converting it into slopes, the landscape takes on a wholly new aspect and the region is said to be *mature* (Fig. 49). The intricate network of drainage is complete and all the inter-stream areas have been carved into slopes. The main streams have cut downward far enough to decrease their own gradients appreciably. As decreasing gradient progressively decreases each stream's downcutting power, and thus causes it to linger at each successive level, the force of the current deflected from side to side of its valley begins to cut effectively and each valley is thereby widened. In this way valley

widening increases proportionately as valley deepening decreases, by the process known as lateral planation. Long stretches of the main streams are at grade as they wander over their newly developed valley flats, while the upland (initial) surface, so prominent during youth, has melted into slopes. Since downcutting has been greatly checked, the mantle of loose material on the valley sides is not swept away as rapidly as it is formed by the agents of weathering. It therefore accumulates, and moves slowly down the slopes under the influence of gravity (locally aided by slope wash, ground water, and frost action). In this way hol-

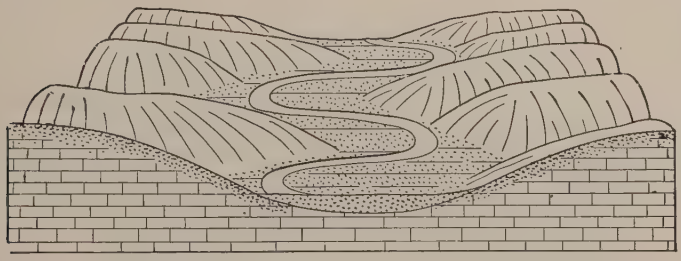


Fig. 54. — The smooth curving profiles of maturity, showing their relation to accumulated mantle rock which masks the horizontal strata. Compare the valley shown in Fig. 34, in which the horizontal strata appear unmasked because of continued rapid downcutting by the stream.

lows in the slopes are filled in, irregularities are smoothed out, and the profiles of the valley sides are converted into smooth flowing curves (Fig. 54). It is for this reason that the period of maturity is the time of restful beauty in a landscape. The rugged splendor of youth has been modeled into sweeping mellow curves.

Old Age Stage. — Because of the low and ever-decreasing gradients of the main streams, the heights are now wasted by the steeper tributaries much more rapidly than the larger valleys can be cut down. The result is a gradual decrease in relief as the divides are worn down to lower and lower levels (Fig. 50). Lateral planation by the main streams widens the valleys and thus helps to cut away the adjacent higher land. The streams are sluggish, meandering widely. Natural levees bordered by swamps are developed in their lower courses. As the bottoms of the main valleys slowly approach baselevel, corrasion by the main streams gradually ceases, and only the divides, where the gradients are still appreciable, are notably lowered. In this way the hills of maturity melt down into low elevations in old age, shedding the run-off feebly in sluggish streams. Erosion takes place more and more slowly. The last few feet of vertical cutting might require a longer time than the entire amount of preceding excavation. The resulting surface of low

relief, very gently undulatory, is called a *peneplain* ("almost a plain," Fig. 51). The highlands have been brought low; the rocks that compose them have been carried bit by bit to the sea and have been there deposited in beds as sediment. Only a few residuals of the former high land remain. Here and there isolated hills (Fig. 51) rise above the general surface, like islands above a sea. They are composed either of very resistant rock or of masses so far from the main streams that they only stubbornly allow themselves to be graded down to the general level of their surroundings. Such island masses are called *monadnocks* after Mount Monadnock in New Hampshire which rises in this manner above the level of the surrounding country.

The vast plain of central Russia has been cited as a good example of a modern peneplain; it has been slightly raised and the rivers have been set at work again eroding. Ancient peneplains, which have been uplifted and then carved by the streams into tracts of hilly country, have been recognized in many places, such as southern New England, Pennsylvania, central Missouri, and the south of England.

It must be clearly understood that the topographic terms *youth*, *maturity*, and *old age* do not refer to periods of years, or to any absolute age. They denote merely stages, defined by the amount of work done in proportion to the total amount of work involved in the cycle. Thus a region of very soft rocks might reach old age while an area of resistant rocks was still in youth as far as the amount of erosion accomplished is concerned. It follows that an extensive valley system might exist in various topographic stages in different localities, depending on supply of water and the varying nature of the underlying rocks. As a matter of fact this is true of the valleys of most large rivers.

Effect of Vegetation on Erosion. — In humid regions the surface is commonly covered with an almost continuous blanket of sod, supplemented locally by brush and forest. This mat of vegetation occupies uplands, slopes, and valley bottoms to the very edges of the streams. Through the action of frost and through repeated saturation with ground water, aided by gravity, the surface soil creeps down the slopes, carrying the vegetation with it. The movement is so slow and imperceptible that the mat of sod is rarely breached. Fresh gullying is hindered for several reasons: (1) Because the mass of roots distributed through the soil, together with the mat of organic matter on the surface, holds the soil firmly together and enables it to resist the pressure of the moving water. (2) Because the mat of vegetation acting like a sponge absorbs the water and permits it to drain off so slowly that the erosive effect of sudden rushes of water after storms is prevented. (3) Likewise in springtime the rapid melting of the snow is hindered by forest shade.

Such effects are of course most noticeable on steep slopes, among hills or mountains. The profiles and contours resulting from this essentially unbroken protective covering, especially in the mature stage of the cycle, are smooth and flowing, a series of beautiful curves (Fig. 289). It is moreover a noticeable fact that in forest-covered countries the flow of the streams is less irregular than in non-forested regions, and the stream waters are relatively clear.

If the forest cover of a country is removed, erosion proceeds rapidly (Fig. 55), and in a variety of ways great damage may be done. The regulative action of the forests on erosion and the flow of rivers is a

Fig. 55. — After the removal of the forest cover the soil has been carried away so rapidly that the remaining trees have their roots exposed by the lowering of the surface. Southern Appalachians. (U. S. Forest Service.)

matter of great importance, not only from the geologic standpoint, but as vitally affecting civilization. In some countries, of which parts of northern China and Spain might be selected as examples, the improvident removal of the entire forest cover has reduced large areas, through displacement and loss of arable soil by erosion, to sterile wastes, subjected alternately to hot and baking droughts and sudden disastrous floods. Destruction of the forests by fire may have a similar effect. Once destroyed, and the soil washed out, they may be restored only with great difficulty after long periods of time. Considerable areas in the southern United States have been much impoverished in this way. In places where density of population places a premium on all arable land, terracing of hill slopes to prevent erosion is much resorted to. The yearly loss of valuable soil is one of the great wastes of modern civilization that should be checked as much as possible; forests should be cultivated on all eminences and places not adapted to agriculture and their cutting carefully governed, not alone for the timber they furnish, but to prevent erosion and regulate the flow of streams.

The Cycle of Stream Erosion in a Semiarid Region

Vegetation and Erosion in Semiarid Regions. — The importance of the vegetation common to humid regions has been outlined in the preceding paragraphs. Let us turn now to drier regions where vegetation is less abundant. In semiarid regions (regions where the annual rainfall is roughly 10 to 20 inches), such as in much of the Great Plains region east of the Rocky Mountains, conditions somewhat resemble those in the deforested areas of the more humid country farther east. Trees are rare, the sod mat is present but not generally strong, and the

Fig. 56. — Steep-walled "wash," or stream channel. The bank is 30 feet high and has been cut since 1880. (Longwell.)

soil is loose, dry, and porous. Rainfall, moreover, is likely to occur in sudden bursts. The result is rapid run-off and hence rapid erosion. Gullying occurs wherever the soil is laid bare, as on cattle trails and in wheel ruts. The streams are subject to sudden and heavy floods, their waters are very muddy, and in months of little rainfall they are commonly low or even dry. Because of the low rate of weathering of their side walls, coupled with rapid downcutting by their streams, the valleys are steep-sided, and in addition they may be flat-bottomed because of excessive deposition (Fig. 56).

Effect of Resistant Rocks; Canyons and Gorges. — During the youthful stage of the cycle in humid regions, the streams cut down much faster than their valleys are widened by weathering and slope wash; and so if the initial elevation of the land is great, canyons and gorges (deep narrow valleys) may develop. But it is equally true that in maturity these must melt into wide, open valleys as downcutting decreases.

In semiarid regions, however, the rate of valley widening by weathering is so slow that canyons and gorges, once they are developed, persist for a longer time. The grandest example is the Grand Canyon in Arizona, one of the most impressive wonders of the world. It is more than 200 miles long, 10 miles wide at the top, and from 3000 to 6000 feet deep. In general its cross section shows a broader upper can-

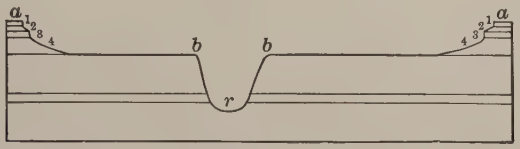


Fig. 57. — Ideal section across the Grand Canyon. (After Dutton.) *aa*, outer canyon walls; *bb*, inner gorge; 1 and 3, hard resistant beds; 2 and 4, soft beds. Vertical scale exaggerated.

yon within which lies a deeper inner gorge (Fig. 58). It is cut in nearly horizontal beds of rock of varying hardness. These rest on crystalline rocks of almost uniform resistance, which in one stretch have themselves been trenched to a depth of 2000 feet in the inner gorge. The more resistant rock layers form gaunt cliffs whose talus slopes partly cover the softer beds, and whose outcrops form broad terraces and platforms. These effects, and the irregular cutting, carving, and recessing of the canyon walls through ravines and side valleys, have given rise to enormous and striking architectural forms (Fig. 58). Some of the masses thus carved out are themselves large mountains. The river is a swift, turbulent stream, heavily laden with silt, from 200 to 300 feet wide and 2400 feet above sea level at the Bright Angel trail, the place in Arizona where the canyon is ordinarily seen. The Colorado must be considered as a young river in respect to the character of its valley and the magnitude of the erosive task that it has yet to accomplish.

Effect of Weak Rocks; Badlands. — When weak materials such as clays and shales are laid open to rain wash, they are rapidly carved into gullies. Striking examples of such erosion are to be seen along the rivers that drain the Great Plains region. These rivers, such as the Missouri and its tributaries, the Cheyenne and the Platte, in places run in valleys sunk a considerable distance below the general level of the country. The rock that forms the sides of the valleys consists for the most part of very soft, barely consolidated clays and sands, easily cut by rain wash and gullying. The result is that on either side of the stream from the bottom-land by the river to the bench-land forming the plain, there lies a gradually rising belt of country dissected in the most intricate fashion by systems of gullies, gulches, and ravines, with spurs, knobs, and sharp ridges separating them. Such tracts of country are

Fig. 58. — Grand Canyon of the Colorado. View is mostly of the inner gorge; the wall of the upper broader canyon is seen in the distance. (U. S. Geol. Surv.)

Fig. 59. — Effect of rain wash in beds of clay. Detail. Sioux Co., Neb. (U. S. Geol. Surv.)

Fig. 60. — A "hoodoo" in Monument Park, Colo. Hard masses of ironstone in beds of soft sandstone have shielded the rock below them from erosion and have thus produced pillars.

known as *badlands*, because of the difficulty experienced in traversing them.

Sculptured Forms: Buttes and Mesas; Hoodoos. — It has been explained that weathering of rocks is not everywhere uniform. All parts are not equally accessible through cracks and fissures by the agencies that produce decay, and some parts may be harder and more resistant than others. Because of this want of uniformity, remnants of the more resistant material are left as projecting masses. These masses protect the softer rock below, and pillars are thus formed (Fig. 60). Where these forms exist on a small scale they are often referred to as "hoodoos." Large isolated masses of soft rock capped and thus protected by hard layers are known in the western United States as *buttes* (Fig. 61). Some are of mountainous size. Very broad flat-topped

Fig. 61. — Red Butte, Bell Ranch, New Mex. (U. S. Geol. Surv.)

features, plateau-like in form, are termed *mesas* (Spanish "table"). Mesas are capped by layers of hard rock, in many cases lava, which have protected the softer layers beneath. These features testify to the great amount of material carried away by erosion from around them. Both buttes and mesas are present in humid regions but their appearance is not usually striking, because weathering and slope-wash conceal their framework.

Summary. — The influences wrought by slight rainfall and consequent scanty vegetation impose themselves on the process of erosion throughout the semiarid cycle. The drainage pattern remains the same as in a humid region save that the tributaries are scanty instead of numerous.

Alluvial fans play a much more prominent part in the semiarid than in the humid landscape, but following their extensive development in the youthful stage they are gradually cut away as the land is lowered, and the resulting surface is a peneplain.

The Cycle of Stream Erosion in an Arid Region

Some authors classify arid regions roughly as regions that receive less than 10 inches of rainfall annually; others classify them as areas in which the drainage does not reach the sea. On the latter basis, one-

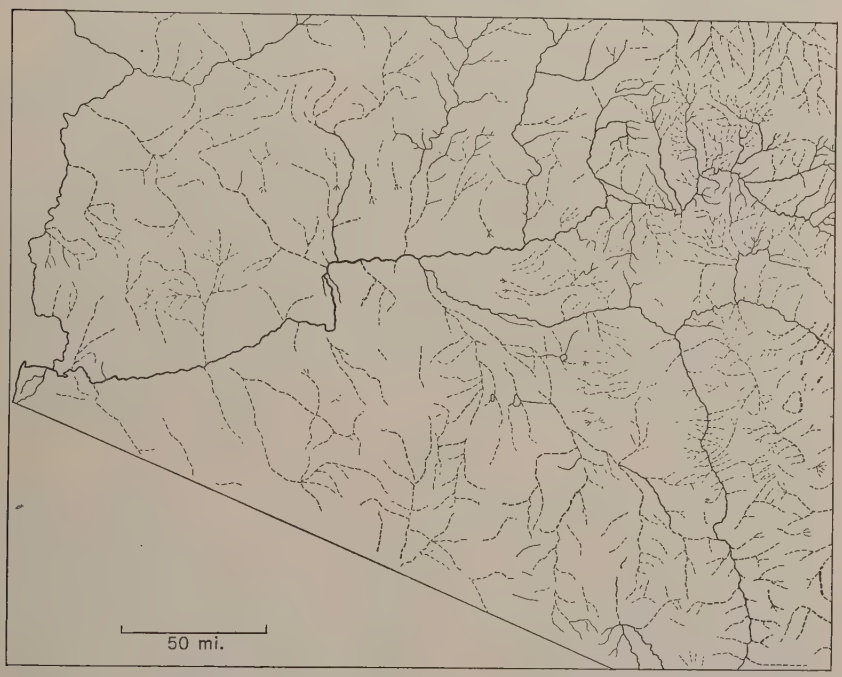


Fig. 62. — Outline map of a portion of southern Arizona showing both through-flowing and interior drainage. The large east–west stream is the Gila River. Many of its potential tributaries never reach it, and most of them (shown by dashed lines) are intermittent in their flow. Compare with the area of similar size shown in Fig. 297

quarter of the land area of the globe is arid, if exception be allowed for through-flowing streams like the Nile and the Colorado. These streams maintain themselves through arid tracts in spite of great evaporation and lack of many tributaries because their headwaters in distant mountains give them a large and steady supply.

Chief among these areas of *interior drainage* are the Sahara, the Libyan Desert, and the Kalahari in Africa, parts of the great Basin and Range

region between the Wasatch Mountains and the Sierra Nevada in the western United States, the desert of western Australia, certain basins high in the Andes, and wide areas in central and western Asia. All are alike in that the streams which drain them lose themselves in the interior (Fig. 62). But they differ in many respects. In some, streams are active agents of erosion and deposition, while in those that receive the least rainfall, the wind seems to be the chief dynamic agent. The latter have not been thoroughly studied. Because of this, and because no such deserts exist in North America, the discussion which follows is based upon the deserts of the United States.

Controlling Principles of Denudation in Arid Regions

Several dynamic processes are important in arid regions. The process of mechanical weathering is universal and is not further discussed. The other processes are as follows:

(*A*) **Stream Erosion and Deposition.** — Rainfall over the desert ranges sends down torrents of water through the dry gullies and gulches and out on to the plains below. The excessive load of loose weathered material acquired in the steep gulches is rapidly deposited in broad alluvial fans as the stream velocities are abruptly checked and as the streams lose volume through evaporation and sinking. Fans at the mouths of adjacent valleys coalesce and in time build up broad apron-like *piedmont plains*. The alluvial fan, comparatively rare in humid regions and common in semiarid regions, is the universal unit of stream deposition in regions of aridity.

(*B*) **Sheetflood Erosion and Deposition.** — In certain districts where cloudbursts are rare but exceptionally violent, the flood water debouching from adjacent mountain valleys spreads out and coalesces on the lower slopes. Moving with a continuous creeping or rolling motion the flood sweeps all before it, the water front, two or three feet high, "curling over and breaking in a belt of foam like the surf on a beach." The transportation and deposition of waste by these *sheetfloods* is an important process even though the floods are comparatively rare.

(*C*) **Mudflows.** — Another normal though infrequently operative agent of gradation in arid regions is the *mudflow*. This is a process intermediate between a sheetflood and a landslide. It occurs only where earthy material becomes watersoaked on steep slopes after heavy rains, and moves downward and outward as a slippery mass (Fig. 63). It advances in waves, stopping when it becomes too viscous to flow and damming the water behind it until it liquefies again and proceeds, like

an advancing flow of lava. Mudflows can carry boulders many feet in diameter. Observers have seen these great rocks bobbing "like corks in a surf." In the course of time, successive mudflows play a large part in erosion and deposition.

Fig. 63. — Margin of a fresh thin mudflow. East side of the Stillwater Range, Nevada. (Blackwelder.)

(*D*) **Landslides.** — Landslides are important in some places where slopes are steep, where rocks are jointed so as to form great heavy blocks of talus, and especially where impervious shales make a slippery base for the mass of débris. The most favorable conditions for landslides are not found in arid regions, but locally they play a part in erosion.

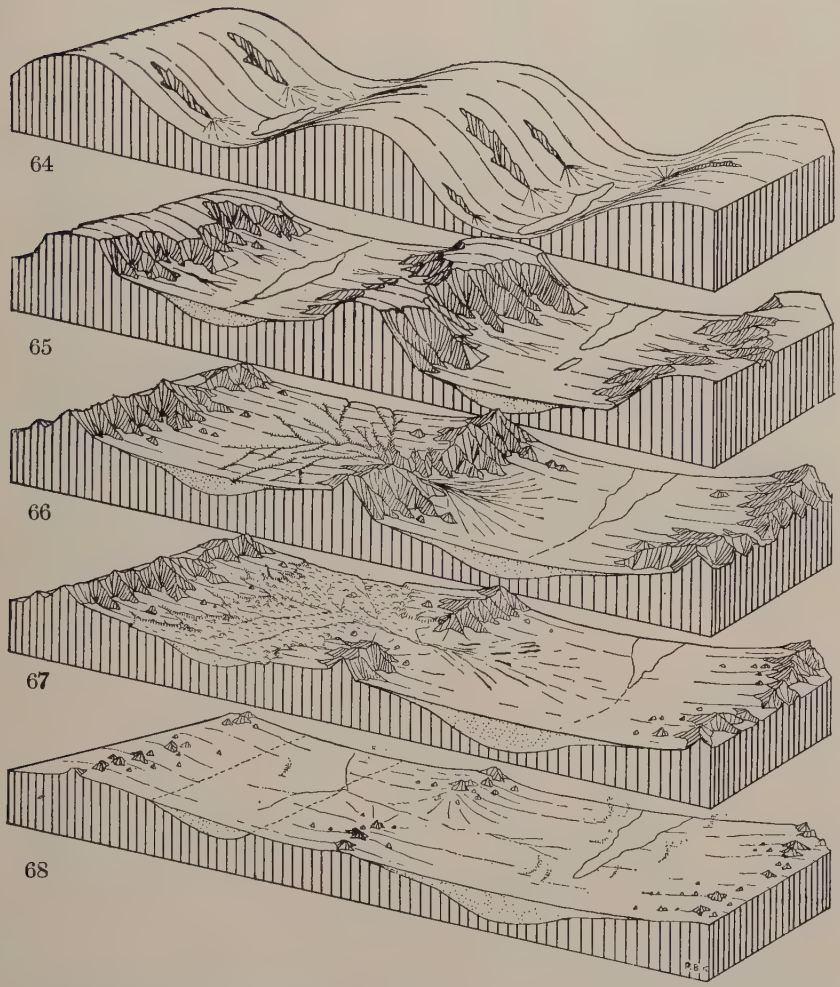
(*E*) **Deflation.** — *Deflation* is the picking up and exporting of fine material by the wind. The rapid mechanical weathering and scanty vegetation in arid regions are factors favorable to deflation; and in some deserts, as in the Sahara where sandstorms are frequent and violent, deflation plays an important rôle in degradation. The numerous sand dunes indicate the temporary resting place of the material in transit, and ships in the South Atlantic testify to the amount of fine material blown into the sea by the prevailing winds. Deflation is not, however, of great importance in the American deserts, its effects being largely masked by the agents cited above.

Outline of the Cycle

(1) **Initial Stage.** — The following brief outline, following the studies of Davis, is organized with reference to Figs. 64–68. The conditions considered are to a certain extent special, but general principles are illustrated. Let a land surface in an arid climate be warped up into sharp folds. With the beginning of uplift, consequent streams develop, carrying drainage from the new highlands down into the adjacent troughs (Fig. 64). They flow only after violent storms and are represented at other times only by dry valleys. Most of them evaporate or sink beneath the surface before they reach the bottom of the trough, depositing their loads in rows of alluvial fans which flank the highlands. Growth and coalescence of the fans result in the narrowing of the intermont basins. The floors of these basins are called *playas* (Fig. 114), and the waters that reach them are impounded as shallow *playa lakes*, only to evaporate soon afterward. The broader term *bolson* (Spanish "purse," i.e., a pocket or basin without outlet) is commonly used to describe the whole waste-filled valley (Fig. 69) between two desert ranges. Figure 64 shows the initial surface, initial consequent valleys, and two playas in the downwarped troughs.

(2) **Youthful Stage.** — Instead of being increased as in the normal cycle, the relief is slowly diminished by the removal of waste from the highlands and its deposition on the lower slopes and in the playas. In this way two bolsons are developed (Fig. 65), their centers occupied by playas and their sides flanked by ragged mountain escarpments strongly dissected by steep gulches and valleys. The baselevel, instead of being single and fixed as in a humid climate, is multiple, being formed by the various bolson surfaces; and is moreover gradually rising as the deposits rise at the expense of the wasting highlands. With decreasing gradients, the mountain streams deposit farther and farther up their valleys. Thus the heads of the fans migrate slowly backward toward the mountain crests. All the streams are intermittent, flowing only after rains. A certain amount of deflation takes place, especially of the fine material on the playa flat.

(3) **Mature Stage.** — As the mountain divide between two bolsons is cut down and the bolson floors are concomitantly built up, the higher basin in time comes to drain downward *across* the old divide into the lower one. When this occurs, the mature stage is said to be reached, and drainage may pass from the upper basin to the lower in times of rain (Fig. 66). The upper basin thereby begins to be dissected by a consequent system of gullies working headward from the new channel, and the débris from their excavation is deposited as a great fan in the



Figs. 64–68. — Ideal cycle of stream erosion under an arid climate. (Compare Figs. 46–51.)

Fig. 64. — Initial stage, showing development of alluvial fans and playas.
Fig. 65. — Youthful stage, showing decrease of relief as the bolsons rise by filling.
Fig. 66. — Mature stage, showing capture of the higher bolson by the lower one.
Fig. 67. — Later mature stage, showing dissection of the higher bolson, transfer of the waste to the lower, and the exposure of pediments.
Fig. 68. — Old age stage, showing disintegration of the drainage, low relief, and climax of wind action.

lower basin, hastening its filling. The surface of the lower basin has now become the master baselevel controlling both its own streams and those of its neighbor.

Full maturity is reached when the upper basin is so thoroughly dissected (usually into badlands, since they are composed of loose uncon-

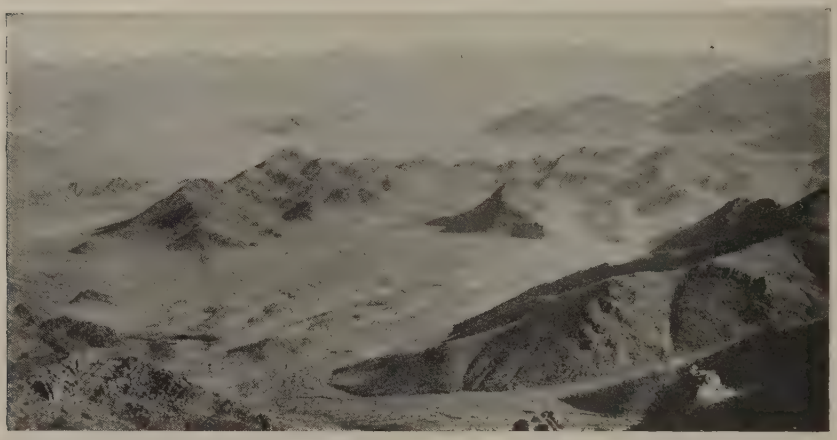


Fig. 69. — Bolson between Desert Range and Sheep Range, Nevada. Mature stage of the cycle. Compare Figs. 66 and 67. (Longwell.)

solidated deposits) that every part of it drains down into the master basin (Fig. 67). The initial highland surface is gone, and the mountains are cloaked ever more completely with waste.

(4) **Old Age Stage.** — With the lowering of the mountains the rains, infrequent at the outset, become even more rare since condensation decreases with decreasing relief. The whole process outlined above is correspondingly retarded, but the wind, less hampered by stream action, tends to erode hollows in the fine loose material of the basins. This breaks up the drainage pattern by forming hollows in various parts of the master bolson. The higher basin, stripped of most of its earlier mantle of waste, is largely floored with bare rock planed to a platform or *pediment* by the debouching mountain streams (Figs. 67 and 68). The wind becomes an increasingly important agent of erosion, disintegrating the drainage still further, and slowly lowering by deflation the whole surface, which now resembles a plain more nearly than a pair of basins. Only the rock masses that are most resistant to weathering remain as monadnocks. Since the wind can work at will over the entire area, the surface is slowly worn down to essentially the same level (Fig. 68). This final floor is thinly veneered with waste and dotted with monadnocks.

This last stage does not exist in any known desert at the present time probably because sufficient time has not elapsed since the recent Glacial Period (apparently a time of almost universal humidity) to allow it to develop. Penck, a German geologist, argues that such a surface could be worn down well below sea level, providing the sea were held out by surrounding highlands. The ultimate baselevel must be the ground water surface (Chapter VI) below which the mantle rock would be moistened and deflation thereby stopped.

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CHAPTER V

GLACIERS AND GLACIATION

Glaciers carry off the accumulated snowfall from the lands, as streams remove the surface waters. They are rivers of ice. Wherever the winter snow does not melt in the summer it accumulates, and where such accumulation continues, glaciers originate. The study of existing glaciers discloses the pronounced changes they effect upon the land, and the observation of similar effects upon lands where today there are no glaciers leads inevitably to the conclusion that glaciers must once have existed there. The present is used as a key to the past. It will be our purpose, then, to study first the existing glaciers; next we shall examine the geologic work they perform, and then reconstruct the glaciers of the past.

The Growth of Glaciers

Perpetual Snow Fields. — On all the continents, with the exception of Australia, there are places where some of the winter snow remains unmelted from year to year, giving rise to perpetual snow fields. Such places are more numerous in high latitudes than in low, and perpetual snow fields are a familiar sight on lofty mountains, even on those that lie beneath the equator. The level above which snow is perpetual, is known as the *snow line* (Fig. 70). At the equator it lies from 15,000 to 18,000 feet above sea level, in Mexico 14,000 feet, in Yellowstone Park 10,000 to 11,000 feet, in southern Canada about 9000 feet, in southern Alaska about 5000 feet, in Greenland about 2000 feet, and in arctic America a few hundred feet. The height of the snow line is determined not only by altitude and latitude (temperature), but also by the annual precipitation, humidity, and location on the sunny or shady side of a mountain. For example, in Bolivia the snow line is 18,500 feet in elevation on the dry western side of the Andes, and 16,000 feet on the moister eastern side, and in parts of Alaska and Siberia where the ground remains frozen throughout the year the mean annual temperature is low enough for perpetual snow fields, if there were sufficient snowfall to exceed wastage by evaporation.

Perpetual snow fields exist on most of the lofty mountain chains of North and South America. They are widespread in the higher mountains of Europe, Asia, and New Zealand; smaller ones occur even in

tropical Africa. The greatest fields of snow and ice are in Antarctica and Greenland.

In southern Alaska, where the winds from the warm Japan current, heavily charged with moisture, rise over the high southern mountains, there is an unusually heavy snowfall, which may amount to 50 or 60

Fig. 70. — Snow line on Mt. Fairweather (15,330 feet), Alaska; from the Pacific Ocean. (Alaska Glacier Studies, Nat. Geog. Soc.)

feet in a year. This is the greatest region of mountain glaciers in the world, both as to total area and the number of individual glaciers.

Névé; Change into Ice. — The greatest accumulations of snow occur in those parts of the perpetual snow fields whose angle of slope is less than that at which snow will slide. They are called *gathering grounds*, or, in the case of mountain snow fields, *catchment basins* (Fig. 71). They receive not only that snow which falls directly on them but also that which slides off the steeper slopes above. Under its own weight the snow becomes compacted and at the same time changes in character. The loose, feathery, newly fallen snow soon assumes a granular texture like coarse sand, and resembles the hailstone type of snow such as we see in the spring, in the remnants of winter snowdrifts. It is called *névé*. The change takes place largely as the result of alternate thawing and freezing at the surface, during which the larger grains of snow grow at the expense of the smaller ones. Whole snow fields so transformed become *névé fields* or slopes.

Beneath the surface, if the thickness of snow is great, the névé becomes compacted as the air is excluded, and passes into dense ice. This ice is more or less distinctly stratified, or banded, due to successive

snowfalls of somewhat different consistencies, or to the presence of wind-blown dust.

Movement; the Glacier Formed. — If snow and ice continued indefinitely to accumulate in the gathering grounds, there would, even in the brief space of historical records, be a thickness on the Alps that would reach to perhaps twice the elevation of Mont Blanc. But before any

Fig. 71. — The gathering grounds of the snows. Snow and névé fields on Mt. McKinley (20,300 feet). Note the transfer of snow by a snowslide from the upper steeper slopes to the catchment basin below. (La Voy.)

considerable thickness can accumulate the masses of ice spread slowly outward and downward. Movement commences, and glaciers are formed. Each glacier, in mountainous regions, is a live river of ice by means of which the excess snow and ice are drained off from the higher places to the lower. It moves slowly down the valley, profoundly changing it on the way, to a place where eventually its front melts.

The exact starting point of a mountain glacier is somewhat vague. Its upper limit is usually considered to be the snow line, but there is movement in the névé fields far above this level. In most glaciers there is a zone of prominent cracks, called the *bergschrund*, between the snow slopes and the glacier.

Not every snow field gives rise to a glacier; some are too small to form more than a tract of névé that passes into ice beneath. Such patches are common in all high mountains, as, for example, in Colorado where the general height of the mountains and the amount of precipitation are not adequate to cause real glaciers.

As an intermediate stage between a snow field and a glacier, some large névé fields form at their lower ends ice masses that give evidence of some movement but do not project as ice tongues for any appreciable distance below the snow line. Such masses are called *glacierets*, *hanging glaciers*, or *cliff glaciers*. Most of the so-called glaciers of the Rocky Mountains and the Sierra Nevada Mountains of the United States are of this class (Fig. 72).

Fig. 72. — A glacieret. Shepard Glacier, Glacier National Park. (Alden.)

Lower Limit of Glaciers. — Glaciers disappear either by melting or by emptying into the sea. The distance to which a mountain glacier descends below the snow line before being halted by melting, depends upon the balance between the forward movement and melting. It might be likened to the distance a rod of ice could be thrust into a furnace before being melted; this depends upon the size of the rod, the rapidity with which it is pushed forward, and the heat of the furnace. Thus, in warm regions, glaciers in general project but a short distance below the snow line; if a glacier is large, or flows with relative rapidity, it will extend farther than under the reverse conditions. As we go to higher latitudes, the glaciers reach lower elevations; in sub-arctic regions they actually reach the sea and break off to form icebergs. The lower limit is also influenced by special climatic conditions, for in moist regions, where there is abundant snowfall, the glaciers are larger, flow more rapidly, and therefore descend farther, than in dry regions at the same latitude.

In the Alps and in Norway, glaciers project as far as 5000 feet below the snow line. In southern Alaska they reach sea level at about 55° N.

latitude, and in southern Greenland at about 60°; whereas in Norway at 70° N. they melt before reaching the sea. In southern New Zealand at latitude 45° S., glaciers descend into tropical forests, and in Chile, glaciers from the Andes reach the sea at about 47° S.

Classes or Types of Glaciers

Glaciers are classified according to their shape, size, and location, as: *valley* or *alpine* glaciers; *piedmont* glaciers; and *ice caps*. Enormous ice caps, called *continental* glaciers, overran the northern countries in the recent geologic past, but are known only indirectly through their effects on the land surface. An understanding of them will be made clearer by considering first the separate types of existing glaciers.

Valley Glaciers. — It is valley glaciers that are usually thought of when glaciers are mentioned, for they are the common kind the world over. They are fed by mountain snow fields and flow down existing

Fig. 73. — Typical valley glacier with branches; the Siachen Glacier, Himalaya Mountains. Moraines of earth are seen on its surface as dark bands.

valleys. They have been likened to rivers because, like rivers, they follow the valley windings and spread from side to side, and commonly they have tributaries (Fig. 73) that join the master ice stream and swell its mass. Furthermore a glacier may split about islands and come together again downstream as an unbroken mass of ice. But the resemblance to rivers is largely superficial; differences are more pronounced than are similarities.

There are about 2000 valley glaciers in the Alps. It was here that they were first studied, and that is why they are often called Alpine glaciers. Most of them are less than 2 miles in length; a few are from 3 to 5 miles long, and one, the Great Aletsch, is nearly 10 miles long. Similar valley glaciers may be seen in other parts of Europe, — in Norway, the Pyrenees, and the Carpathians. Magnificent glaciers, up to 30 miles in length, are to be found in the Himalayas, and other high ranges of Asia, except the Altai Mountains, furnish fine examples. Even under the equatorial sun of Africa there is a small glacier on the

slopes of Kilimanjaro (20,000 feet). In the United States hanging or cliff glaciers are numerous, but true glaciers lie only on some of the lofty volcanic peaks in the west, such as Mounts Rainier, Shasta (Fig. 74), Hood, and Baker. Those on Mount Rainier, in Washington, attain a length of 7 miles. The coastal mountains of British Columbia contain many fine glaciers. But the grandest glacial region of the world is in southern Alaska, where valley glaciers exist in unknown thousands, or

Fig. 74. — Mt. Shasta and Shastina from the west. Note that long glaciers descend on the north side, but none on the south side. (U. S. Army Air Corps.)

perhaps tens of thousands, nourished by great unexplored snow fields that mantle the lofty mountains. Some attain 50 miles in length and 5 to 6 miles in width, and descend from heights as great as 18,000 feet above sea level. Scores of them enter the sea and give rise to icebergs.

Piedmont Glaciers. — The name implies a glacier at the foot of a mountain. It designates those glaciers that have descended from the mountains and have spread bulb-like upon the gentler sloping plains beneath. The expanded foot of the Rhone Glacier (Fig. 81) may be considered as the beginning of a piedmont glacier. Usually several glaciers coalesce to form a piedmont glacier. If a valley glacier be likened to a river, then a piedmont glacier may be compared in size to a

lake. Such glaciers are not common, and are confined to high latitudes; the Malaspina and Bering glaciers of Alaska are examples.

The great Malaspina Glacier lies at the base of the range that supports Mt. St. Elias (18,000 feet) and Mt. Logan (19,540 feet), and spreads over an area of 1500 square miles to the edge of the sea (Fig. 75). It has a gently rolling surface, broken by innumerable fissures, irregular hummocks, and piles of débris. Its borders are so mantled by dirt that

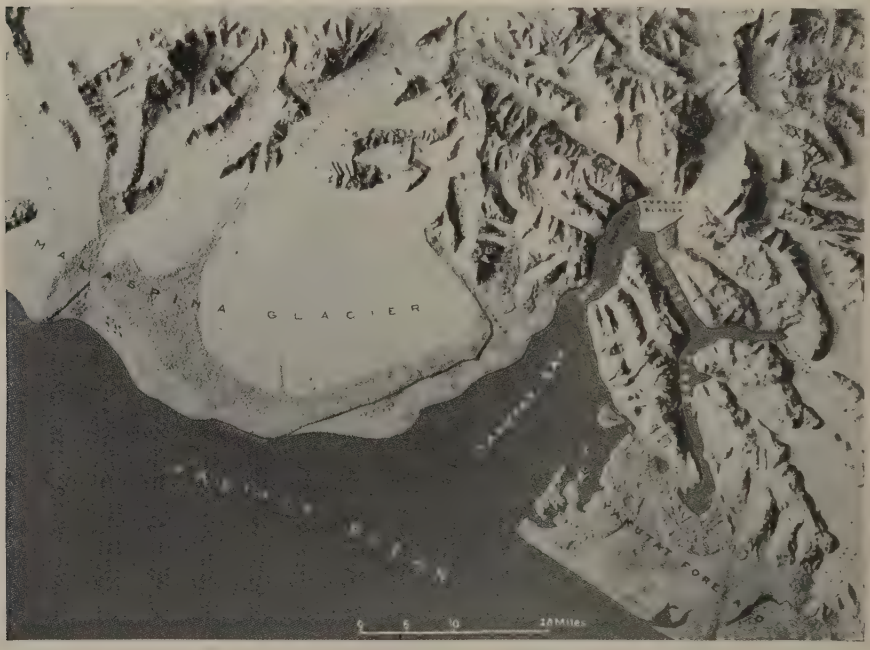


Fig. 75. — Model of Malaspina Glacier, Alaska. (Martin, Univ. of Wisconsin.)

they support a dense forest which rests on 1000 to 1500 feet of ice. The melting ice nourishes several short rivers of large volume.

Ice Caps. — These are vast sheets of ice which, to carry the analogy further, resemble seas in size. The picture they present is that of an almost endless monotony of desolate, wind-swept ice, gently rolling, and with occasional mountains of rock, known as *nunataks*, projecting through them like islands. At their borders they taper down to elongate lobes, or to tongues of ice that discharge into the ocean, forming icebergs.

Only two large ice caps exist today, in Greenland and Antarctica; smaller ones occur in Iceland. The Greenland ice cap is about 1300 miles long and has an area of about 715,000 square miles. It has now been crossed several times notably by the explorer Koch. He states

that there are two great flattish domes from which ice spreads out in all directions. The northern dome reaches an altitude of nearly 10,000 feet and the southern about 8500 feet. The thickness of the ice probably ranges from 2000 to 7000 feet. The edge of the ice is definitely known to be in motion locally, since it discharges into the sea. In the northern part of Greenland, however, where the snowfall is light, the ice is thin and stagnant.

The Antarctic ice cap is thought to have an area of 5,000,000 square miles. Its interior has been partly explored by Shackleton, Amundsen, Scott, Byrd, and Wilkins. Its thickness is unknown but its surface reaches an altitude of about 12,000 feet. According to Scott, it pushes off the land out over the sea to form vast stretches of floating fields of ice, known as the "Great Ice Barrier." The huge tabular icebergs of the Antarctic break off from the barrier ice.

The great continental glaciers of the Ice Age, which have now entirely disappeared, must have resembled the ice caps of Greenland and Antarctica.

The Movement of Glaciers

That glaciers actually move had long been suspected but was not generally known until 1827, when Hugi built a hut on the Aar Glacier, and its change of position was observed. Since then many accurate measurements have shown not only the rate but also the nature of the movement. The movement is always downstream, even though a glacier is said to be retreating; a retreat of a glacier is of course not a bodily movement of the ice back towards its source, but simply a retreat in the position of the ice front owing to excess of melting over forward motion.

Rate of Movement. — Hugi's hut moved down the Aar Glacier a distance of 4650 feet in 15 years; 44 years later it was found 7900 feet down the valley, indicating a rate of movement of from 6 to 10 inches per day. A similar long-time measurement came to light for the Glacier des Bossons on Mont Blanc. In 1820 three guides were buried beneath an avalanche on the mountain. It was predicted by Dr. Forbes in 1858 that their bodies would be given up by the glacier 35 to 40 years after their burial. Just 41 years later the heads of the three guides, with some hands and clothing, appeared at the foot of the glacier, so well preserved that they were recognized by friends. The average rate of movement was 8 inches a day. Other measurements have been made by placing markers on the ice and noting the time they have taken to travel a measured distance. Such measurements in the Alps have shown that the rate of movement there seldom exceeds 1 to 2 feet per day. In Alaska measurements made on the Kennecott Glacier (25 miles long;

1 to 4 miles wide) showed an average rate of movement of the central portion, over a period of 1 year, of $2\frac{1}{2}$ inches per day; whereas the Childs Glacier in 1916 moved 4 feet per day, and in 1910, 8 to 40 feet per day. The large Muir Glacier was found to have a motion, at its center, of 7 feet per day. Rates as high as 60 to 75 feet per day have been recorded in some of the tongues of ice that move toward the sea through narrow mountain passes from the Greenland ice cap.

The rate of movement is influenced by several factors: it is greatest in those glaciers that have a large supply of ice; it increases with a steep valley slope and a smooth bed; it increases with a steep upper slope of the ice; it is greater in summer, when there is more water in the ice, than in winter; it increases when the temperature of the ice approaches the freezing point. The gradient, the amount and thickness of ice, the steepness of the upper slope, and the temperature, are thus the chief factors that affect the rate of movement.

Differential Movement. — An important discovery in regard to glacier motion is that the rate of movement is not the same in all parts of its mass. A glacier does not move as a whole by sliding down its bed, like a cake of ice off the roof of a house. It has been found that stakes driven in a straight row across the top of a glacier after a time become curved downstream, proving that the middle portion moves faster than the sides. Similarly, vertical lines of pegs on the side of a glacier show that the top moves faster than the bottom. Also, it has been observed that the line of swiftest motion does not always lie in the middle of the glacier, but is sinuous. The movement of the ice is therefore *differential;* that is, some parts of it move faster than other parts. This conclusion is important to a proper understanding of the cause of glacier motion.

Nature of Glacier Movement. — The observation that glacier movement is differential led early observers to assume that the flow of glacier ice is similar to that of a stiff, viscous fluid, such as pitch or asphalt. This, however, cannot be the case; the viscosity is only apparent, and not real. In a fluid, the molecules are free to move readily in any direction as they do in water or pitch, but in a crystalline solid the molecules or atoms are fixed in definite positions from which they can be moved only by great force. Ice is a true solid; its component particles are definitely arranged with respect to each other in a geometrical pattern. Since it is crystalline, it cannot flow by free movement of the molecules and therefore cannot exhibit the property of a fluid. Furthermore, a moving glacier may contain many fissures and since flowing fluids do not crack, ice cannot move as a liquid. Also, a glacier will not turn and flow up an empty tributary valley, even though the top

of the ice may tower high above the tributary floor; the Kennecott Glacier forms a dam several hundred feet high across Hidden Creek tributary, but the ice does not flow into Hidden Creek valley. Were the ice a true fluid, it would spread out into the empty tributary valley. Ice must flow as a solid; it does so because it becomes plastic under stress, as does a ductile metal, and so resembles a viscous substance.

Explanation of Glacier Movement. — The mechanism of glacier movement is complex and several ideas have been advanced to account for it; but to discuss them in full is beyond the scope of this work. Certain factors that have been considered important are: (1) Melting of ice under pressure and refreezing when the pressure is removed; (2) deformation of ice crystals along certain gliding planes without destroying its crystalline structure; (3) rotation of granules of ice; (4) interchange of individual molecules between ice granules; (5) shearing or sliding of one mass of ice over another. No one of these factors by itself is thought to be the sole cause of glacier motion, but rather that two or more of them operate together to produce movement. For convenience, these individual factors may be discussed separately.

Melting and Refreezing. — When the water from warm-weather melting at the top of the glacier descends into the ice where the temperature is lower, and there freezes, the expansion upon freezing causes a push in the glacier which will be in the direction of least resistance, or down the valley; also during its travel before refreezing, the water tends to move down the valley as well as down into the glacier, thereby causing a transfer of some of the bulk of the glacier to a lower place.

Melting and refreezing have also been considered to aid movement by allowing individual granules of ice to rotate down grade; under pressure, the granules will liquefy where they bear on each other; the minute films of water so formed will move short distances to places of less pressure, where they will refreeze. This is called *regelation.* Thus the films of water move, and the granules from which they came may rotate slightly in response to gravity. Minute movements are set up, and if similar movements are taking place throughout all the glacier, there may be a slow motion of the glacier down the valley.

Exchange of Molecules. — Up in the névé fields, snow changes into granules of crystalline ice, and the larger of these grow at the expense of the smaller. There is thought to be an interchange of molecules from one granule to the other, in which the molecules move from a place of higher to one of lower pressure, and this would be down the valley. Such molecular interchanges taking place throughout the whole glacier may give rise to a movement of the whole glacier.

Shearing. — It has been noticed, particularly in the lower ends of glaciers where the ice granules are large and interlocked, that there are planes along which one part of the ice mass has sheared over a lower part. The rate of slipping of upper masses over lower masses has actually been measured. It has been thought that this slipping or shearing accounts for part of the movement of glaciers, particularly in the lower ends.

Crystal Gliding. — In some crystalline solids of which ice is one, the component particles (molecules or atoms) are less tightly held together in certain directions than in others. Under stress, movement takes place along such planes of weakness between layers of molecules, without destroying the crystalline character. It has been demonstrated that such *gliding*, as it is called, takes place in ice along certain planes (for example, parallel to the base of the hexagonal prism), and ice, therefore, can adjust itself to pressure by movement as a solid. If such gliding should occur wherever ice granules happen to come into the proper position, a movement may be imparted to the whole ice mass.

Probably several of the factors mentioned above play a part in bringing about motion. In view of the observation that movement is faster in spring than in fall and winter, and increases when the temperature is near the freezing point, it seems probable that liquefaction and regelation, aided by molecular interchange, play an important part in inducing movement. However, it is well known that many cold crystalline metals also flow, when stressed, under conditions of temperature where melting and refreezing are excluded. One need only deform a piece of lead or copper to be aware how small a force is required to make these weak metals flow. Ice is also a weak solid, and the mechanism of its flow may be similar to that of other weak crystalline solids. Recent investigations with the flowage of metals prove that, as in ice crystals, gliding also takes place in metal crystals, and that in addition granulation of the individual grains enables movement to take place.

Various Features of Existing Glaciers

The surface of a glacier must not be thought of as an expanse of smooth ice; typically it is highly irregular, so that travel over it may be laborious and hazardous. It is traversed by wide and deep cracks called *crevasses*, and is covered in places by accumulated heaps of earth and rock fragments called *moraines*. Unequal melting also gives rise to a variety of relief forms that hinder the traveler.

Crevasses. — Among the most striking features of a glacier are the great yawning cracks or crevasses in its upper surface. They may be

20 feet or more in width and up to 300 feet in depth, and may occur anywhere between the névé fields and the lower end. They are formed by tension acting on the brittle ice, and are induced where the ice rides over an uneven bed, or where one part moves faster than another part. Their abundance varies with the amount and the rate of the straining.

Fig. 76. — Ice fall of the Franz Josef Glacier. New Zealand. Almer Glacier entering on left. (New Zealand Geol. Surv.)

They are most prominent where a glacier flows over a sharp incline in its bed, giving rise to an *ice fall*, as in Fig. 76. *Transverse* crevasses usually bow downstream owing to the faster movement of the glacier in its central part. Even a change of slope in the bed of only two or three degrees produces crevasses, a striking proof that ice is not a true viscous substance. Where a glacier emerges from a constricted to a wider part of a valley, *longitudinal* crevasses occur, because the glacier tends to spread sideways and the ice is pulled apart. These are particularly

characteristic of the terminal lobe of a glacier, as in Fig. 81. *Marginal* crevasses commonly occur along the sides of a glacier. These usually point inward and upstream at an angle of about 45° to the course of the glacier. They are formed as a result of tension set up by the central part moving faster than the marginal parts.

Crevasses, once formed, do not remain open indefinitely. Like the particles of water of a river that break over a fall and join again below, so the ice breaks over an irregularity in its bed and heals again into a solid mass by refreezing; all evidence of crevasses may be obliterated. Not uncommonly, however, the upper parts of crevasses are widened by melting so that slices of ice between parallel cracks become sharpened to thin blades and needles, called *seracs*. These may persist long after the lower parts of the crevasses are closed, and a considerable part of the glacier may present a maze of sharp ridges such as one meets in trying to cross the Mer de Glace. On large valley glaciers, such as those of Alaska, these ridges may take the form of steep hillocks 100 to 200 feet high.

Fig. 77. — Slide rock descending from bluffs upon a small glacier (white foreground) to form a marginal moraine. The mine buildings are situated upon the edge of the marginal moraine. Kennecott, Alaska. (Bateman.)

Moraines. — In a river valley, the débris that is worn from its sides forms local talus slopes or is washed down into the stream and borne away. But in a valley occupied by a glacier, this material descends upon the ice to form a marginal fringe of débris (Fig. 77) that is carried along slowly by the moving ice. Thus there are formed continuous bands of débris along the margins of the glacier; these are called *lateral moraines* (Fig. 78). Where tributary glaciers enter, *medial moraines* are formed (Fig. 78). Glaciers may be seen with several medial ribbons of débris, each separated from the others by "white ice," and the number present usually indicates the number of tributaries that have joined the master glacier. As the glacier reaches lower levels and undergoes surface melting, the ice beneath the débris melts less rapidly, so that commonly the moraines may be prominent

ridges up to 100 feet in height. The débris may slide down the sides of these ridges and so widen the morainal bands.

The débris is not only on top of the ice (*superglacial*) but is also within the body of the glacier (*englacial*). Much of the latter is carried along near the bottom of the glacier. The englacial material is derived partly from débris that fell on the surface as the glacier was being built up; partly from material that fell into the crevasses; and partly from material that was scraped off the sides and bottom of the valley. The englacial material may become superglacial where rapid melting occurs near the end of the glacier, and, in combination with spread-out lateral and medial moraines, may cover the entire "snout" of the glacier. The transported material that is eventually dumped at the end of the glacier, in a confused mass of earth and stones, forms the *terminal moraine.*

Fig. 78. — Plan of a valley glacier, showing tributaries and terminal lobe; *ll*, lateral moraines running into *tt*, the terminal moraine; *mm*, medial moraines. As melting progresses, more and more material appears; *s*, exit of subglacial stream; *vv*, valley train of water-laid débris.

Glaciers are thus great transporters of débris; fine dust, and boulders of thousands of tons weight are carried with equal facility.

Differential Melting. — The presence of débris on the surface of a glacier causes uneven melting by the sun, and numerous irregularities result. Thin layers of dark débris or dust on the glacier will absorb more heat from the sun than does the adjacent white ice, and will become warmed through and melt the ice beneath to form *dust wells* or "bath tubs." If the débris layers are thick, they act as heat insulators; the ice beneath is protected from melting and prominent ridges or *dirt cones* are formed. Some large slabs of rock eventually cap ice pedestals (Fig. 79) and are known as *glacier tables.*

Drainage. — If one walks over a glacier on a warm day, he hears the murmur of innumerable rivulets of water that are formed from the active melting that goes on. Many join together to make streams so wide that they are difficult to jump across, and these may be seen to plunge down deep crevasses, commonly into circular shafts called *moulins* or *mills*, of their own making. And deep beneath, one may hear the subdued roar of larger subglacial streams that flow in ice tunnels

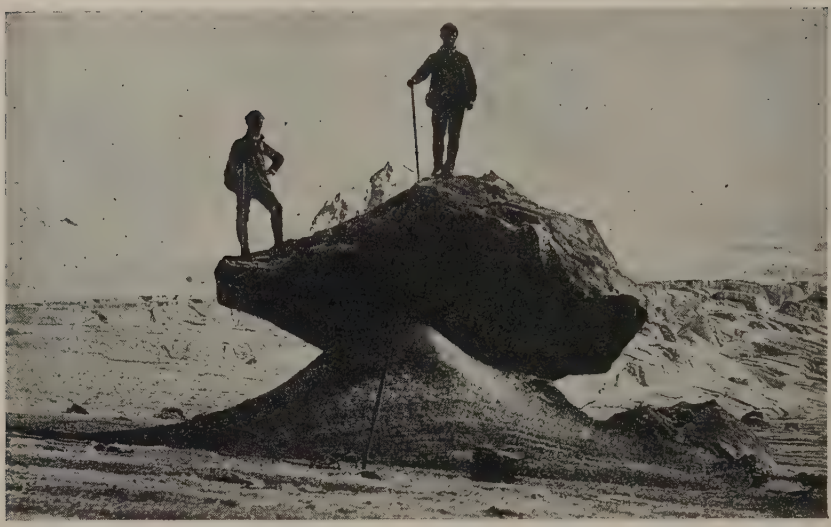


Fig. 79. — Glacier table mounted on ice pedestal; caused by differential melting. Switzerland.

Fig. 80. — Subglacial stream and ice cave. Morainal material above falls as the ice melts and helps to build the terminal moraine. Transported blocks fill the bed of the turbulent stream which carries away the finer earth and ground-up rock. Chamonix, Switzerland.

that they have made. All along the sides, also, numerous trickles of ice water descend and commonly form marginal streams. In a warm summer many feet may be melted from the top of a glacier. On one small glacier in Alaska, telephone poles sunk 6 feet in the ice had to be reset twice during the summer.

The subglacial streams emerge from caves at the front of the glacier (Fig. 80), or less commonly they spurt forth from large vertical holes and form the starting points of good sized rivers. The water is striking in appearance, for it has a peculiar opaque grayish-white color that has given rise to the name *glacier milk*. This is due to the presence of fine unweathered rock material that has been ground from the glacier bed and is carried in suspension. This milky water may be traced far down the valley or out to sea, and its presence makes the traveler suspect that he is approaching a region of melting glaciers.

Fig. 81. — Two views of the Rhone Glacier, Switzerland; the upper one taken in 1870, the lower in 1905, illustrate the retreat of the ice in 35 years. The older one shows the terminal lobe and longitudinal crevasses.

Advance and Recession of Living Glaciers. — Glaciers appear to advance and retreat in response to varying climatic cycles of greater or less precipitation, or of sunshine and cloudiness. At present there seems to be a general condition of recession, although there are at the same time minor oscillations of retreat and advance. In Europe and North America glaciers have been retreating, in general, for several decades. The Muir Glacier in Alaska has retreated more than 8 miles in the

last 25 years, and since 1887 the Illecillewaet Glacier in British Columbia has retreated at least 500 feet. The amount of retreat of the Rhone Glacier is shown in Fig. 81. In the Chamonix Valley in the Alps, however, the glacial retreat from 1812 to 1910 amounted at most to only $\frac{1}{2}$ mile.

On the other hand, some individual glaciers have shown prominent advances. For example, all the glaciers on the Savoy side of the Mont Blanc chain advanced prominently between 1910 and 1920. The Grindelwald Glacier, following a long retreat between 1875 and 1900, had by 1924 advanced a considerable distance. In Alaska, a lobe of the Childs Glacier advanced 2000 feet in 1910 to 1911, receded for the next few years, and advanced a short distance again in 1925. These advances appear to be only temporary oscillations, although in the case of the Alaskan glaciers the length of time over which observations have extended is too short to state definitely whether there is a general retreat or advance.

GEOLOGICAL WORK OF GLACIERS

Studies of living glaciers teach us that they erode, transport, and deposit materials. And, with the work of existing glaciers in mind, we are led to infer, when similar effects are observed elsewhere, that glaciers were widespread in parts of the northern and southern hemispheres where today there are none. This was an ancient *continental glaciation.* Existing glaciers and their work are to be seen chiefly in mountainous regions, but the results of continental glaciation are nearly everywhere present in northern regions, even in flattish areas. The work of the vanished continental glaciers can be interpreted properly only by an understanding of what has been accomplished by living glaciers, or the present remnants of them, in the mountainous regions. Consequently, our attention will first be directed to glacial work in mountainous regions, which we can see actually taking place or which has taken place so recently that to fill in the gap involves but little speculation.

Erosion by Mountain Glaciers

Valley glaciers by their erosion not only give rise to distinctive floor features and characteristic valleys, but also profoundly modify those parts of mountains that project above the glaciers. How these features are produced will become clear by an understanding of the nature of glacial erosion.

Nature of Glacial Erosion. — Glaciers erode by *plucking* and by *abrasion.* The first process is most active in the upper reaches of a valley

glacier, but operates throughout its course; abrasion is effective wherever the glacier moves in a well-defined channel.

In plucking, the glacier actually quarries out masses of rock, incorporates them into itself, and carries them along. This is accomplished by water that trickles down into crevices in the rock, freezes, and springs out blocks of rock, or that freezes around blocks bounded by joints or crevices so that they become a part of the body of the glacier, or by the ice of the glacier being pressed into the cracks in the rock by its own

Fig. 82. — A group of glacial cirques (amphitheaters); also a small valley glacier, and a glacial stream and lake. Coast Range, British Columbia. (Bateman.)

weight. Then, as the glacier moves forward, the blocks are plucked out and dragged along with it. It is a powerful quarrying operation. This process is most effective at the edges of the névé slopes, where the *bergschrund* facilitates ingress of water by day and freezing by night. Here there is a constant quarrying inward and downward, and in the course of time this gives rise to bowl-shaped basins called *amphitheaters* or *cirques* (Fig. 82). Commonly these are cut somewhat deeper at the center than at the place of discharge, and when the glacier disappears, these depressions become filled with water (Fig. 83) and form some of the beautiful mountain lakes or tarns, such as Lake Louise in the Canadian Rockies. Such cirques are common features of high moun-

tains in middle and high latitudes, and are evidence of the former existence of glaciers.

The process of plucking operates also on the sides and bottoms of a valley glacier, particularly where projections of rock extend into the ice and where the bedrock is much jointed.

Dirt and stones frozen fast in the bottom and sides of the ice form a huge slow-moving rasp that grinds away or *abrades* the bedrock of the bottom and sides of the valley. As the cutting tools wear out, new ones are continually supplied by plucking and abrasion. This rasping power is enormously augmented by the great weight of the overlying ice.

Fig. 83. — Lion Lake and small tarn in glacial cirque south of Triple Divide peak, Sierra Nevada, Cal. (Matthes.)

The fact that a glacier erodes the sides of the valley it occupies, no less than its bottom, makes its erosional work entirely different from that of a river. The shape of a glacial valley is thus quite different from that fashioned by a river, and is distinctive.

The rate of erosion is influenced by various factors. It is obvious that the larger the glacier, the greater will be its erosive power, because of its great weight; consequently master glaciers erode their valleys more than do smaller tributary glaciers. A thin ice sheet may override soil without even removing it. Also, clear ice produces no abrasion; the greater the amount of débris in the ice, up to certain limits, the more it can grind the bedrock. Erosion is favored by fast movement of the glacier and by weak bedrock. Another factor is the firmness with which the teeth of the rasp are gripped by the ice; where the ice is soft and melting, as at the lower ends, it may not even remove loose soil. Plucking plays a small part if the bedrock is smooth and un-

jointed, but becomes highly effective if the rock surface is rough and jointed.

In contrast to a river, a glacier can cut far below grade locally and scoop deep hollows in bedrock along its course. As a result, glaciers erode valley bottoms unevenly; pronounced deepenings of a glaciated valley occur in constricted parts, where the ice has been forced to move faster, or in places underlain by exceptionally weak rock.

Results of Erosion. — Results of glacial erosion are evident in bedrock features, in the character of the valleys, and in superglacial effects.

The Valley Floor. — If one walks up a valley recently vacated by a glacier, he will notice that the bottom and sides are smoothed, rounded,

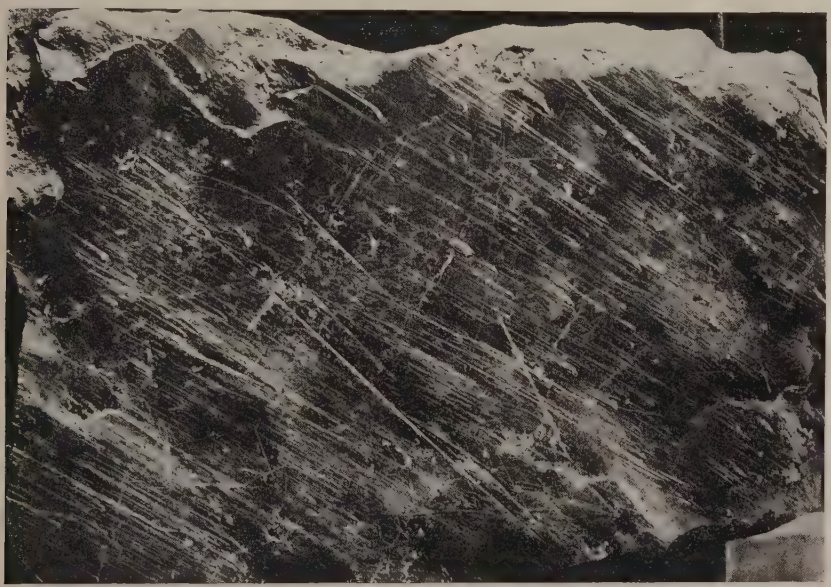


Fig. 84. — Glacial striae; scratches and groovings made by moving ice on limestone bedrock. Near Rochester, N. Y.

and scored by scratches and grooves, called *glacial striae*, that trend in the direction of glacial flow (Fig. 84). They were caused by the rock-shod bottom ice that dragged heavily over the floor; the pebbles and boulders scratched and grooved, and the dust polished. The rock flour thus formed was swept away by the subglacial drainage and gave rise to the "glacier milk" previously referred to. As one ascends farther, he may have difficulty in climbing steep little steps that face him abruptly but taper off smoothly upstream. These result from the plucking of blocks from the downstream sides of jointed hummocks of

bedrock. The traveler may have to circle around many ponds or pools that occupy gouged-out rock basins in the valley floor. If he glances down the valley, he may note that the floor is made up of smoothed, rounded rock masses, elongated in the direction of flow, which, from their resemblance to the backs of a flock of sheep, are called *roches moutonnées* (Fig. 85). Some valleys also have a prominent stepped profile to which Russell has given the apt term " a cyclopean stairway." The presence of all these features is a clear indication of the former existence of a

Fig. 85. — Roches moutonnées near Mono Pass, Sierra Nevada, Calif. (Matthes.)

glacier. But they are not confined to mountain glaciers alone, for similar features have also been produced by continental glaciers.

The Glacial Valley. — The shape of the valley is also a distinctive feature of erosion by mountain glaciers. It has a characteristic U-shape, flaring at the top, with steep sides, and a broad bottom. (Fig. 86.) Commonly it is several thousand feet deep. No river erosion could produce such a valley. A comparison of Figs. 86 and 88 (*a*) shows the contrast between a glacial valley and a normal youthful river valley. This unusual shape results from the fact that a glacier erodes the sides as well as the bottom of its valley, and cuts away the ends of interlocking spurs developed between the tributaries of a stream-eroded valley.

The tributaries also are peculiar. They do not join the master stream at grade as they normally do in river valleys, but commonly enter hundreds, or even a thousand or more feet, above the main valley floor. The streams that flow along such *hanging valleys*, as they are called, when they reach the abrupt drop, cascade down to the main valley floor and

give rise to some of the most beautiful waterfalls (Fig. 87), such as those of the Yosemite Valley in California, the Lauterbrunnen in Switzerland, or the Yoho Valley in British Columbia. The hanging valleys originate because the large master glaciers erode their valleys more rapidly than do the smaller tributary glaciers. Consequently, the master valley becomes overdeepened by bottom erosion with respect to the tributary valley and the latter is left hanging, or the master valley becomes overwidened by lateral erosion, causing the mouth of the tributary to retreat upstream and to be left hanging above the bottom of the main valley. Both processes operate simultaneously, but in the case of high hanging valleys, vertical cutting has probably been much more effective than lateral cutting. Even where hanging valleys are not present there commonly exists an abrupt change in slope between the steep glacial valley wall and the unglaciated country above. In Switzerland these "shoulders," as they are called, are usually 1000 feet or so above the valley bottoms and are a favorite place for villages. In Alaska they reach heights of 3000 feet or more above the valley floors.

Fig. 86. — U-shaped glacial valley. Hodnett Lakes Valley, Yukon Territory. (Geol. Surv. of Canada.)

Fig. 87. — A hanging valley and falls. Yoho Valley, British Columbia.

Other distinctive features of glacial valleys are truncated or *faceted* spurs. They are spurs of normal river valley erosion that have had their "noses" cut off by the erosion of the valley glaciers. Their characteristic triangular form may be seen

in Fig. 88, (*b*) and (*c*). Thus U-shaped sections, cirques, rock basins, hanging tributary valleys, shoulders, and facetted spurs are characteristic erosional features of glaciated valleys. The changes effected in a normal river valley by glaciation are seen in Fig. 88.

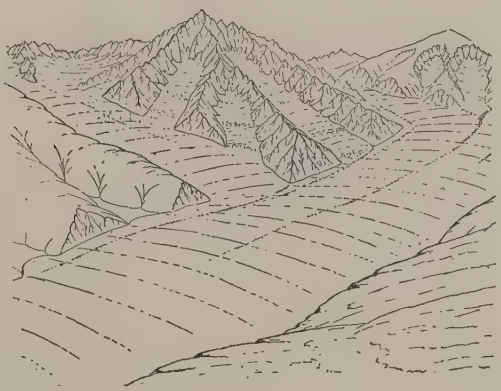
Fiords. — Where overdeepening in glacial valleys along a coastal region has extended beneath sea level, or where the coast has been depressed, the sea occupies the valleys after the ice has vanished, and *fiords* result. Most of the fiords of Alaska, British Columbia, Labrador, Chile, and Norway probably have been formed in this way. They are essentially no different from any glacial valley except that their bottoms are covered by the sea instead of being occupied by a river. Depths of water of more than 5000 feet have been measured in Alaskan fiords, and of 4000 feet in the Sogne Fiord in Norway. The depth of the Alaskan fiords is thought to be due to the depression of the coast more than to submarine excavation.

Superglacial Erosion. — The higher parts of mountains that rise above the glaciers and are covered by perpetual snow-fields are also greatly affected by ice erosion. This is shown graphically in Fig. 88, (*a*) to (*c*). Parallel tributary glaciers in the process of widening their valley sap the sides of intervening ridges and leave jagged *comb ridges* or *arêtes*. The headward gnawing in cirque formation causes cirques to approach each other from opposite sides of a ridge until they intersect, leaving jagged sawtooth pinnacles between them (Fig. 82). Or, three or four cirques disposed about a single mountain and eating headward, as on the right hand mountains in Fig. 88 (*b*), will coalesce until an irregular or pyramidal *horn* is all that remains of a once massive mountain. Such horns are typical of glaciated mountains; the famous Matterhorn in the Alps and Mount Assiniboine in the Canadian Rockies are outstanding examples. The erosive work above the glaciers is thus chiefly sapping and undermining, and the results are ragged pinnacled slopes whose precipitous character usually defies the mountain climber.

Deposition by Mountain Glaciers

The vast amount of material transported by glaciers must be deposited when the glaciers melt. Consequently different kinds of deposits are formed, depending upon the character of the material deposited, the manner of the deposition, and the shapes and positions of the resulting depositional features. The materials transported and deposited by a glacier are called glacial deposits, or *glacial drift*. But the drift is of two distinct varieties: — one, which is dumped in heterogeneous fashion as the ice melts, and is unstratified, is called *till*, or *boulder clay;*

(*a*) A mountain mass, normally eroded by weathering and running water and unaffected by glacial action. The valleys and ravines are V-shaped in section.



(*b*) The same mass, strongly affected by glaciers which occupy the valleys. Note the rugged topography above the ice, produced by weathering and frost.

(*c*) The same mass after the retreat and melting of the ice. Note the nature of the topography, the trough-like form of the glaciated valleys, the amphitheaters, some with lakes, the hanging valleys, and the faceted spurs. (W. M. Davis.)

Fig. 88.

the other is material that is sorted out and stratified by water from the melting ice, and laid down to form *glacio-fluvial* deposits.

There is much similarity in the deposits of mountain and continental glaciers. Consequently the descriptions given for mountain glacier deposits will apply also in part to those formed from continental glaciers.

Unstratified Deposits

The outstanding characteristic of glacial till is that it is unstratified and in this respect it differs from the nicely assorted water- and wind-laid sediments. Coarse boulders and fine sediments occur jumbled together just as they were dropped by the melting ice. Some of the pebbles and boulders are as unworn and angular as when they were removed from their original positions. Those that were dragged against the bedrock on the bottom or side of the valley, however, have characteristic smooth flat surfaces or *facets* (Fig. 89) and are polished, striated or scratched. No other pebbles or boulders are quite like them; they are diagnostic of glaciation.

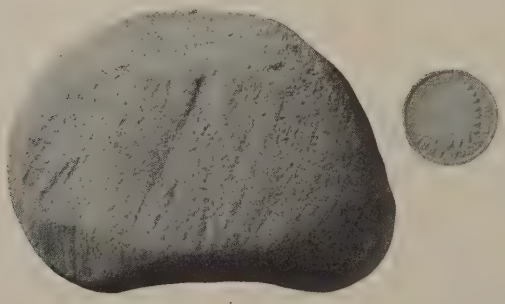


Fig. 89. — Facetted and striated pebble. Sierra Nevada, Calif. (W. D. Johnson.)

Moraines. — When a glacier melts, the débris contained in the ice is dropped as a till sheet or *ground moraine* over the valley bottom. Commonly it is not very thick, and much of the finer material may be washed away, leaving the boulders as the most conspicuous part of the moraine.

The lateral moraines of the glacier are left along the sides of the valley as long ridges of till, some of which are several hundred feet in height. The name *lateral moraines*, used for these features on the sides of live glaciers, is retained for the ridges after the glacier has disappeared. In places the ridges are sharp-topped, because the material fell to its angle of repose as the ice melted. The medial moraines of the glacier less commonly persist as distinct medial ridges on the ground; they are likely to be washed away, owing to their central position in the valley. In favored places, however, parts of old medial moraines are well preserved.

The *terminal moraine* consists of jumbled masses of till, usually of

irregular shape, extending across the valley. Commonly it has a general crescentic form, concave toward the glacier. It is composed of pebbles and boulders, facetted and angular, heterogeneously admixed with earthy materials. The piles range in height from a few tens to a few hundreds of feet. The higher ones usually indicate that the ice front was relatively stationary for a long time so that much débris was deposited at one place. A glacier that retreats comparatively fast leaves only a low, scattered terminal moraine. If the ice front in its retreat makes extended halts at intervals, a series of *recessional moraines* will be left. The terminal moraine is usually breached by streams and may be entirely removed by them. The surface of a typical terminal moraine is characterized by numerous rounded mounds and small undrained basins, with a haphazard distribution. This peculiar kind of surface, known as *knob-and-basin* topography, is caused by highly irregular deposition of the terminal moraine. This irregularity results from frequent minor fluctuations of the ice front and from variation in amount of débris delivered by the ice to any one point. Some of the basins in an abandoned terminal moraine form small ponds.

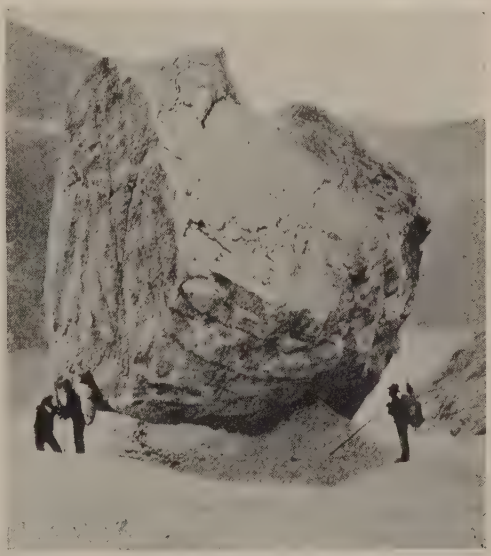


Fig. 90. — Erratic block, Alaska boundary. (Coleman, after Melson.)

Glacial Boulders, or Erratics. — Another characteristic feature of a glaciated valley is the presence of scattered boulders, or *erratics*, of all shapes and sizes, many of which are foreign to the underlying rock; boulders of granite derived from the head of the valley may rest on limestone in the lower part of the valley. Commonly they are perched in insecure positions (*perched boulders*), or they may be so nicely poised that they can be rocked by the hand (*rocking stones*). They may be distinguished from somewhat similar boulders of decomposition by their dissimilarity in composition from the underlying rock. These erratics will be referred to again under continental glaciation. One may be seen in Fig. 90.

Glacio-fluvial Deposits or Stratified Drift

The streams discharged by melting glaciers carry out boulders, pebbles, sand, and silt. The boulders, however, are quickly dropped; the remaining materials build up an *outwash plain*, or a *valley train* which may extend far down the valley. Nearest the glacier are the coarser pebbles, and farthest away are the sands or coarser silt; the finest rock flour of the milky water may be carried far beyond the valley train. The deposits are thus sorted out by the water and are stratified

Fig. 91. — Pitted outwash plain, derived from Hidden Glacier, Alaska. Part of ice near front of glacier shows in foreground. Note the braided streams, which are building up the plain. (U. S. Geol. Surv.)

into beds of sands and pebbles. The latter are somewhat rounded by the friction they experience as they are rolled along. Across these gravelly plains, where the gradient is low, the streams pursue a braided course and are continually shifting their channels. They thus work over the material and search out the finer sands to carry them farther downstream, leaving behind extensive beds of well-sorted, clean pebbles, whose thickness may be scores of feet. As the outwash material accumulates it may lap up over the end of a stagnant glacier, or bury large isolated blocks of ice. Later, when the ice melts, pits or *kettle holes* result, and the stratified deposits form a *pitted outwash plain.* (Fig. 91.)

CONTINENTAL GLACIATION

Continental glaciation on a far grander scale than that of mountain glaciation has taken place over the greater part of high-latitude lands. The glaciers themselves have largely vanished; their former presence is inferred from the results they achieved, and their character is arrived at by inference and by analogy with existing ice caps. The nature of the evidence, however, in the light of our present knowledge is conclusive. They overran regions where mountains are few and small, as well as some mountainous districts; and they are supposed to have been great flat domes of ice, so thick that they covered most of the lands. Their action, therefore, was different from that of valley glaciers; they did not flow as tongues of concentrated erosive power in valleys, and therefore glacial valleys, hanging valleys, and cirques are not characteristic of their work; they flowed from centers that were hundreds of miles from their terminals, and hence carried some materials great distances; they carried little or no débris on their surfaces, and as they had no side boundaries they had no lateral moraines. Their morainal material came largely from the underlying bedrock, and therefore much of it shows signs of wear; their melting took place over a wide front, with the result that their terminal moraines and outwash deposits are of vast extent; they overrode hill and vale, and hence much of their deposit is independent of topography. We shall first examine the signs they have left and then deduce their character.

Continental Glacial Erosion

The Floor. — The continental glaciers were powerful engines of erosion; over wide areas they removed all soil and weathered rock, and plucked and abraded the floor, giving rise to vast stretches of gently rolling surfaces that display glacial striae, groovings, and roches moutonnées on fresh unweathered rock. These roches moutonnées have a smooth, gently sloping up-glacier side and a more abrupt down-glacier slope. The glaciers overrode hills and sculptured them into shapes similar to the smaller roches moutonnées. But toward the southern edge of the ice advance, pronounced erosional features are less conspicuous, although striae and grooves are abundant; in places the weathered surface rock has not been entirely removed. Erosional features in the regions of continental glaciation are, on the whole, less evident than depositional features.

Glacial Lakes. — Throughout the glaciated region of eastern North America are hundreds of thousands of lakes, large and small, that have

resulted from glaciation. In the northern areas they have been formed chiefly by the flooding of rock basins scooped out by the ice. In the southern regions, a more important cause is the formation of depressions by irregularly deposited débris, or the deposition of moraines athwart valleys, forming dams that pond the drainage. So numerous and closely spaced are these glacial lakes that what would otherwise be a little known wilderness in northeastern North America is made accessible to travel by canoe. In Minnesota alone there are 10,000 lakes. Of the many beautiful lakes that lend charm to the scenery of the rolling country of New England, the Adirondacks, Minnesota, Canada, or Norway, by far the greatest number have resulted from the work of glaciers. In the unglaciated country of the southern and central United States, lakes are rare or wanting.

Transportation by Continental Glaciers

The continental glaciers carried great quantities of rock débris which later was dropped on the land or spread afar by the action of water. Because of their great size and distribution, they transported vastly greater amounts of débris than valley glaciers. Most of the material of the drift has not been moved more than 50 miles, although some of it has travelled hundreds of miles. This has been determined by the finding of materials far distant from their place of origin. For example, slabs of native copper that came from the great copper lodes of the Keweenawan Peninsula of Michigan have been found as far south as Missouri; likewise, boulders of an unusual conglomerate containing reddish pebbles of jasper found throughout Ohio have been traced to a bed on the north shore of Georgian Bay, Canada. In northern Sweden, glacial boulders of copper ore that had been transported by the ice were traced back in the direction of ice flow and resulted in the discovery, by electrical surveys in 1925, of Sweden's most important copper mine. Of unusual interest in this connection was the finding of several isolated diamonds of good quality in glacial deposits in Wisconsin, Michigan, Ohio, and Indiana. Their position in the deposits indicated clearly that they had been transported and deposited by the ice, but their source is as yet unknown. It is presumed to be somewhere to the northward in Canada.

Deposits of Continental Glaciers

The disappearance of the continental glaciers left the vast regions overridden by them mantled with glacial débris of various kinds and arrangements. These deposits have exerted a profound effect on the

social and economic development of the countries with a glacial past, for the fertility or sterility of the soil, the drainage, and in part the topography, have been conditioned by such deposits.

As in the case of the existing glaciers previously discussed, deposits were formed of unsorted material dropped from the melting ice, and also by means of running water. Therefore both glacial till and glaciofluvial deposits are widespread in the glaciated countries.

The Till or Unstratified Deposits

The Till Sheet. — When the ice melted, its load of débris was dropped as a sheet of till that overspread the land to a variable thickness ranging up to several hundred feet. It consists of old soil, ground-up rock, small fragments, pebbles, and boulders in heterogeneous disorder, and commonly the adjacent boulders have no similarity in composition. (Fig. 92.) Many of the pebbles and boulders are polished, scratched, and

Fig. 92. — Drumlin near Newark, N. Y. (U. S. Geol. Surv.)

facetted (Fig. 89). Glacial boulders are so common that one may pick them up at random in the glaciated regions; the many stone fences of New England attest their abundance. In places they are so numerous as to render the soil untillable. It is this bouldery character of the till that has given rise to the term *boulder clay*. The till sheet was deposited somewhat evenly over the bedrock, but variations in thickness impart to the surface a gently rolling character, with low *swells* and shallow *swales* or depressions. In some localities there are smooth, oval-shaped hills of till, elongated in the direction of ice movement. These are

drumlins; it is not known definitely how they assumed their form and position (Fig. 92). They are especially common in New York, eastern Massachusetts, and eastern Wisconsin. In general, they average about a half-mile in length and 100 to 150 feet in height; and commonly they occur in clusters.

Moraines. — Near the margins of continental glaciation one may see belts a few hundred feet to several miles broad that look as though an army of excavators had dug a maze of depressions and stacked the débris in nearby dump piles. There is bewildering confusion in the haphazard

Fig. 93. — Glacial till, Bangor, Pa., consisting of unassorted boulders and fine materials. (Pa. Geol. Survey.)

arrangement of pits and mounds. They are spaced without order. Examination of the material itself reveals glacial pebbles and the lack of sorting characteristic of till (Fig. 93). This is the work of the glacier, and these features are typical of its terminal moraine. In its extensiveness, it bears little resemblance to the small terminal moraines of valley glaciers, for these large ones may be followed almost without break across states and countries. However, the peculiar *knob-and-basin* topography in the great moraines has the same general origin as in the analogous valley moraines; it is due to the irregular deposition of débris liberated from a fluctuating ice front. If the ice, for a period of time, melts about as rapidly as it advances, much material will be dropped in one place. As the ice front oscillates, and projecting lobes and embayments change their position and shape, the frontal depo-

sition is necessarily hummocky. The knobs may rise to 100 or 200 feet above the basins, or there may be only a few feet difference in elevation. The depressions are commonly occupied by marshes or ponds that are numbered in the thousands. In places the terminal moraine consists only of a prominent ridge or a series of separate or overlapping ridges.

Every considerable halt of the ice front in its retreat gave rise to other moraines, called *recessional moraines*, and the unsteady retreat of the vanishing glaciers can be traced as surely as though they had been under observation. Long lobes of ice must have lagged behind the retreating main sheet, for in many places the terminal and recessional moraines extend parallel to the direction of movement, indicating deposition along the sides of such tongues. Kettle holes formed from the melting of residual blocks of ice are numerous in the vicinity of the terminal moraine.

Erratics. — Unlike the glacial boulders or erratics of mountain glaciers that are found only in valleys, those dropped by continental glaciers

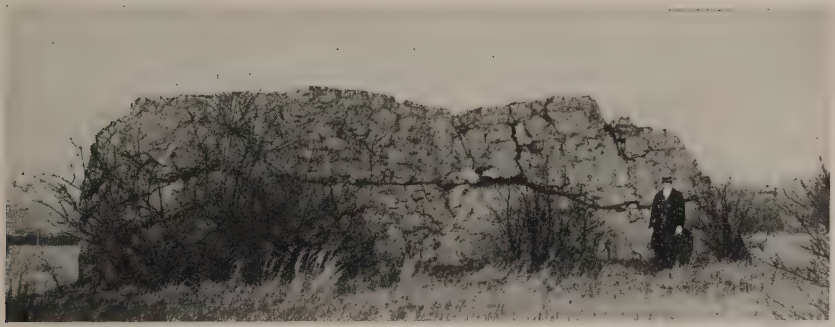


Fig. 94. — Glacial erratic, a transported boulder of trap resting on sandstone; weight about 500 tons. New Haven, Conn.

occur far and wide over hill and vale. It is not uncommon to see them on the very tops of high hills, in positions where they could not have fallen from higher ground or have been carried by streams. And when, in addition, they rest on till and differ in composition from the underlying rock, it is certain they were left by glaciers (Fig. 94).

Glacio-Fluvial Deposits

The disgorging of torrents laden with sediments, as described for the small living glaciers, must also have taken place on a vastly greater scale from the hundreds of miles of front of a continental glacier. This is not mere supposition, for the extensive glacio-fluvial deposits that lie

within and around the regions of continental glaciation abundantly prove it. However, a greater variety of water-laid deposits emanated from continental glaciers than from valley glaciers, and there is evidence that some were formed under the ice, and others near its edge, in addition to those that were laid down beyond the ice front.

Outwash Plains. — In front of the continental glaciers, valley trains such as those described in connection with existing valley glaciers formed

Fig. 95. — Stratified drift. Excavation for Yale Library, New Haven, Conn. (Longwell.)

only where well-defined valleys were present. But valleys were infrequent in the open country fronting the continental ice caps, and most of the water's burden was deposited to form great coalescing alluvial sheets, known as *outwash plains* or *frontal aprons*. Like the smaller valley trains, they are composed or water-sorted gravels and sands, built up to considerable depths by streams that rapidly became graded and braided. Some of the finer rock flour supplied the material to make the loess deposits of the Mississippi Valley and in parts of Europe. Southern Long Island was built up as a sandy outwash plain, as was also a large part of southern Ohio. With the retreat of the ice, outwash plains came to overlie earlier till deposits (Fig. 96).

Some of the outwash plains are prominently pitted by large and small

kettle holes, where the outwash materials have slumped after the melting of stagnant ice masses that were partly or entirely buried by the accumulating sediments. Many of these kettles, some as deep as 100 feet, occur in the sandy outwash plain at New Haven, Connecticut.

The sands and gravels of outwash plains are widely used as materials for road and building construction.

Fig. 96. — Water-laid glacial drift on unassorted till. Columbus, Ohio.

Kames and Eskers. — These are peculiar forms of deposits made by the sediment-laden streams from the melting ice. *Kames* are hummocky hills and ridges composed of stratified drift, which generally occur in clusters with marshy depressions between them. They are thought to have been formed at the glacier's edge, where steep cones of water-laid gravel became heaped up against the ice front, and, as the ice front melted, slumped down as irregular hillocks; or under the frontal ice by gravel-laden streams that flowed into depressions or holes; or in shallow basins of water on top of the ice, from which position the stratified gravels slumped into piles as the ice melted.

Eskers are long, winding ridges of stratified gravel or sand that look like railroad embankments. They are only a few feet wide at the top, and range from 10 to 100 feet in height, but some of them may be traced for many miles in length. They are striking features of the landscape in Scandinavia, Finland, Ireland, and the northeastern United States (Fig. 97). They are supposed to have been formed by deposition from streams that flowed in ice tunnels in and beneath stagnant ice. As the stream beds became built up by deposition of gravels, the streams enlarged their tunnels by melting the roofs. Thus there accumulated long, sinuous deposits of water-laid gravels enclosed by ice. When the

ice melted, the gravels slid down to their angle of repose, to form eskers which trend in the general direction of ice movement.

Fig. 97. — The esker of Punkaharju, Puruvesi, Finland. In Scandinavia such a ridge is called an "ose" (plural "osar").

Lake Deposits. — Remarkably banded clays have been formed in patches within the glaciated regions, notably in Scandinavia and North America. The Swedish geologists call them *varved clays*. In a fresh excavation it is evident that they are horizontal and that most of the individual layers are an inch or less in thickness. (See Part II, Fig. 229.) But the observer is impressed chiefly with the striking uniformity in thickness of the separate bands and the sharp line of demarcation between layers. If he examines a band closely, he will see that slightly coarser sediment at the bottom grades upward into material so fine that it will not grit between the teeth, and that this alternation is repeated in the individual bands throughout the depth of the deposit. Such fine rhythmic banding surely could have been formed only in bodies of quiet standing water, such as lakes, and the repeated alternations of coarse and fine material suggest at once a seasonal deposition. And that, apparently, is the manner in which they were formed. The lakes were temporary features, ponded by morainal material or in part by stagnant ice, beyond the melting and retreating ice front. Into them

flowed graded glacial streams carrying rock flour. The coarser material was deposited during spring and summer, when the maximum flow of water entered the lake; the finest material, at the top of each varve, represents the slow settling during fall and winter, while the lakes and streams were frozen over most of the time. Thus each layer, grading from coarser to finer material, represents one year's accumulation. Professor De Geer, in Sweden, realizing this, counted the annual layers and reached the conclusion that Stockholm was under ice 9000 years ago; that the ice sheet began to leave northern Germany 17,000 years ago, and southwest Sweden 12,000 years ago. He also found that it took about 5000 years for the ice to retreat to a point 270 miles northwest of Stockholm. Similar counting in Ontario by Antevs showed that it has been 13,500 years since the ice receded from north of Georgian Bay.

Marginal Glacial Lakes

The retreating ice cap had along its front places where the ground sloped toward it, and naturally in such sites water accumulated to form lakes. These disappeared when the impounding ice melted. Lakes of this origin, some large, some small, abounded along the ice fronts; the beaches, outlets, and deposits are still visible. One of the most noteworthy of these, known as *Glacial Lake Agassiz*, covered a large area in North Dakota, Minnesota, and southern Canada. It was 700 miles long by 250 miles wide. The only remnants now left of it are the present Lake Winnipeg, Lake Manitoba, Lake of the Woods, and other smaller water bodies. Its extensive deposits constitute the fertile wheat lands of that region. Similarly, a series of lakes of changing outline occupied the area of the Great Lakes and St. Lawrence Basin and extended across eastern Ontario. Their outlet was for part of the time through the Mississippi River, later through the Mohawk Valley to the Hudson River, and finally through the St. Lawrence. Abundant evidences of these old lakes are to be found in well-preserved beaches and lake deposits in the vicinity of the present Great Lakes.

The Glaciers of the Ice Age

In the light of our knowledge of the behavior and results of existing glaciers, the features just described under "continental glaciation" lead inevitably to the conclusion that extensive ice caps once overspread great continental areas during the Glacial Period. The evidence is now so plain and convincing that it is difficult to realize that the idea was first proposed in 1837 by Louis Agassiz, and that a quarter of a century elapsed before the glacial theory was generally accepted. The

older ideas that the débris was transported by the Deluge, or by floating icebergs moved by the Deluge, were dispelled slowly.

If one were to travel widely over the glaciated lands, measure the directions of the glacial striae, roches moutonnées, eskers, drumlins and terminal moraines, and plot them in their proper place on a map, he would find that the glaciers did not originate at the poles. In North America there were three main centers, from which the ice spread outward in all directions. In Europe the main ice cap centered in Scandinavia and spread into Russia, Germany, Holland, and the North Sea. The thickness of this ice sheet is thought to have been 6000 to 7000 feet in Scandinavia and about 1500 feet in the Harz Mountains. The Labrador sheet was at least 6000 feet thick in northern New England, and may have been thicker farther north.

At the same time that the great ice caps were spreading over the lands, mountain glaciation was also extensive; and while the ice caps of America and Europe disappeared, the mountain glaciers receded to their present-day smaller proportions. About four million square miles of America and two million square miles of Europe were glaciated, and many mountain areas in addition. There still remain another six million square miles covered by ice in Greenland and Antarctica, so that during the Glacial Period about one-fifth of the land area of the globe was ice-ridden.

Extended studies of the deposits of the Glacial Period show that there was not a simple advance and retreat of each ice cap. Glacial history is far more complicated. In North America there were at least four, and perhaps five, separate advances and retreats of the ice. These are recognized by later till sheets that cover earlier drift, or that overlie interglacial soils, and by other criteria. There was a widespread retreat of the ice in each interglacial epoch, although it is not certain that the ice disappeared entirely preceding each readvance.

The recession of the ice caps left the lands greatly changed. Not only were they eroded, covered by deposits, and dotted by lakes, but the preglacial drainage systems were profoundly altered. New streams originated, old channels became buried, directions of flow became reversed, and many waterfalls were formed. Deranged drainage on a large scale is evident in the Ohio River basin and in parts of New England.

But what caused these great continental ice caps? The answer to this question has puzzled investigators ever since the fact of continental glaciation became firmly established, but as yet there is no generally accepted explanation. It is becoming clear, however, that they probably did not owe their origin to a single cause but to a combination of causes. The various hypotheses advanced are discussed in Part II

of this book. Here, it need only be mentioned that the most important factors in the origin of continental ice caps seem to be variations in amount of solar energy received from the sun — which in turn affect the storminess on the Earth and the amount of heat stored in the oceans — and changes in the shape and elevations of continents and mountain ranges.

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CHAPTER VI

SUBSURFACE WATER

OCCURRENCE, MOVEMENTS, AND RECOVERY

Water on the surface of the Earth wears away the highlands and fills the depressions with sediment. It is a leveling agency that tends in the long course of time to make the land smoother. But in addition to the work of leveling at the surface, water performs other geological work of great importance, chiefly chemical, beneath the surface. This chapter discusses first the nature, position, and motions of the subsurface water, and the forces that control it, then the work that it performs.

Source and History of Subsurface Water. — A part of the subsurface water ascends directly from the reservoirs of fluid rock within the Earth; another part is water that was trapped in the sedimentary strata at the time the sediment was deposited; but most of it is that part of the rainfall that sinks into the soil and into the bedrock below. This water, once it has penetrated into the ground, may find its way back to the surface as springs and join the run-off; it may be drawn to the surface by capillary attraction through the pores in the soil and be evaporated; it may be sucked up by plants and be evaporated through their leaves; it may find its way to the sea through underground channels without returning to the surface; it may be held for indefinite periods within the pores of the rocks; or it may form combinations with the molecules of certain minerals and so become fixed in the rocks.

Porosity of Soils and Rocks. — The porosity of a material is its property of containing open spaces or interstices and may be expressed as the proportion of the total volume of the material not occupied by solid matter. This property varies greatly in different materials but the existence of subsurface water depends upon the fact that all rocks and soils are more or less porous.

It has been shown that a group of spheres of equal size placed together so that each one touches all those that surround it leave unfilled 25.95 per cent to 47.64 per cent of the total space that they occupy. The unfilled space is pore space and the difference in amount depends upon the manner in which the spheres are grouped and not at all upon their size. Ordinary sedimentary matter is not composed of perfect spheres but of irregularly shaped, more or less rounded grains of varying

size. If these grains are well sorted so that those of one size are collected together to the exclusion of other sizes, the porosity of the material approaches and, because of the irregular shape of the grains, may even exceed the maximum for spherical grains. Poorly sorted sedimentary materials have a much lower porosity because fine grains fill the interstices between the larger ones and reduce the amount of open space.

The porosity of materials of the same general kind varies considerably but the following table gives a good idea of average amounts. These figures show that soils have the highest porosity in spite of the poor sorting of their constituents. This is explained by the poorly compacted condition of most soils. They are kept loose and aerated by worms, ants, burrowing animals, and vegetation as well as by plowing and agriculture in general.

Unconsolidated Material		Consolidated Material (Sedimentary Rocks)		Igneous and Metamorphic Rocks
Sand and gravel	35 per cent	Sandstone	15 per cent	Average porosity less than one per cent
Clay	45 "	Shale	4 "	
Soil	55 "	Limestone	5 "	

Igneous, metamorphic, and well consolidated sedimentary rocks commonly have a porosity much higher than that given because systems of fractures develop within them. Limestone may be rendered very

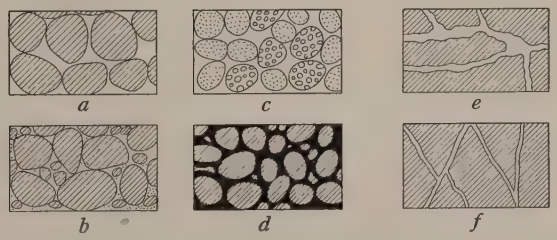

Fig. 98. — Diagrams to show relation of rock texture to porosity. *a*, well-assorted sedimentary deposit having high porosity; *b*, poorly-assorted sedimentary deposit having low porosity; *c*, well-assorted sedimentary deposit made of pebbles which are themselves porous, giving the deposit as a whole very high porosity; *d*, well-assorted sedimentary deposit whose porosity has been diminished by deposition of mineral matter in the interstices; *e*, rock rendered porous by solution; *f*, rock rendered porous by fracturing. (U. S. Geol. Surv.)

porous by the development of cavities due to solution. Figure 98 and the accompanying description illustrates the relation between rock texture and porosity.

Aquifer. — The term *aquifer* is applied to a layer or other body of loose sediment or consolidated rock that yields water in considerable quantities.

Forces that Control Water in Rocks. — Gravity and molecular attraction are the two forces that control the movement of subsurface water. Gravity causes the water to percolate to lower levels or to move laterally down a grade. It also causes the water to issue from springs and flowing wells, forced out by the weight of the water behind. If all the interstices in rocks were large, gravity would be the only controlling force. But, since the actual interstices in many rocks are minute, the molecular attraction of the rock substance acts across the openings and holds the molecules of water firmly in place in spite of their tendency to move downward. This force is called *adhesion.*

Water-Yielding Capacity of Rocks. — Not all of the water contained in the pore space of a rock will be yielded to the natural drainage or to such artificial openings as wells. It has been shown that the size of the grain of a sediment does not affect its porosity but the last paragraph explains how that size does determine the percentage of the contained water that the sediment will yield. A porous and thoroughly saturated clay may be *impervious* to water at usual pressures because the water already in it is held firmly by the molecular attraction of the particles of clay. On the other hand a compact, granular rock such as a granite, with a few scattered fractures, is much less porous but yields the water that it does contain more readily. It is more *pervious* than the clay.

The Situation of Subsurface Water

The water in the ground fills all the pore space up to a certain definite level called the *water table.* The zone below the water table is known as the *zone of saturation* and the water contained in it is called *ground water.* The region between the water table and the surface of the ground is called the *zone of aëration* and the water in it *suspended subsurface water* (Fig. 99). The water table is not a level surface but rather a subdued replica of the land surface beneath which it lies. It is arched up under the hills, sinks a little below the valleys and intersects the surface in some low lying tracts to form lakes, swamps, or springs.

The level at which the water table stands in any particular region depends on the rainfall and the topography. It rises or falls in response to wet or dry seasons and stands within a few feet of the surface in humid regions though it may be several hundred feet in depth in those that are perennially dry. If climatic conditions are the same the water table will stand farther below the hill tops in a rugged region than in one less rugged because the deeper valleys depress the water table near them and give the whole a greater gradient.

The Zone of Aëration. — The thickness of the zone of aëration varies directly with the depth of the water table. It may be nonexistent or

it may be several hundred feet thick, though ordinarily it varies between a few feet and a few tens of feet. It is the zone in which suspended water, present in small quantities only, is held in place despite the downward pull of gravity. This zone is subdivided into three parts: (1) *The soil water zone*, the zone in which water is retained by the minute interstices of the soil and is thus available for plant growth. The

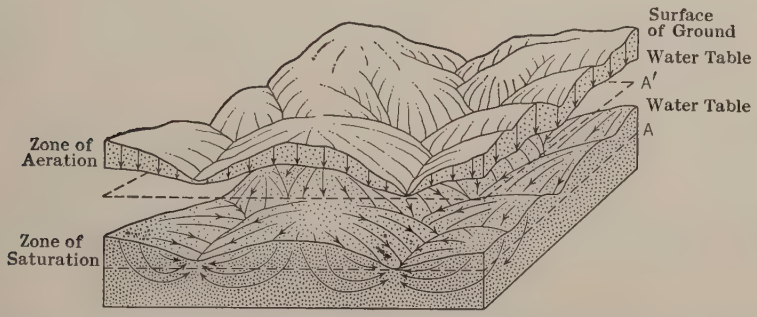

Fig. 99. — Block diagram to show the relation of the water table to the surface of the ground; the direction of movement of subsurface water (indicated by arrows); and the zones of aëration and of saturation.

The upper part of the diagram shows the surface of the ground and the downward motion of the subsurface water as far as the water table. The lower part of the diagram shows the surface of the water table — a subdued replica of the ground surface — and the motions of the ground water in the zone of saturation. Notice that the ground water moves both down the slopes of the water table and down into the ground, and tends to converge at and below the main drainage channels of the region.

The two parts of the diagram are separated in order to show the water table and the movement of the water along it clearly. The dotted lines A and A' would coincide if the two blocks were in proper position.

The subdivisions of the zone of aëration are given in the text.

For the sake of simplification the region illustrated is regarded as underlain by a single stratum of uniform porosity. The difference in the density of the dots represents the different amounts of water contained.

quality of a soil that allows it to hold water in this manner is its specific retention. A poorly sorted, compact soil has a high specific retention while a surface layer of material that yields water readily, such as pure sand, has very low retention. (2) *The capillary fringe*, the zone immediately above the water table that contains water drawn up by the capillary openings in the rock or soil. This zone is thick if the interstices are minute, as in clay or loam, and unimportant if they are coarse, as in gravel or sand. (3) *The intermediate zone* normally lies between the capillary fringe and the soil water but may be absent where the water table is near the surface. It is the zone in which a small amount of the water that percolates from the surface down to the zone of saturation is retained by molecular attraction.

The Lower Limit of the Subsurface Water. — Deep down within the crust of the Earth the weight of overlying matter exceeds the crushing strength of rocks. At such depths the interstices must cease to exist and consequently subsurface water is excluded. This theoretical depth is impossible to state accurately but it lies a number of miles beneath the surface. Actually ground water becomes very scarce after a few thousand feet of rock have been penetrated. Many mines are dry 2000 or 3000 feet down except where deep fissures are encountered. Some aquifers have been encountered at depths as great as 6000 feet, but as a rule wells drilled more than 2000 feet produce little water.

Ground water must be regarded then as a great body of water filling the interstices in all soils, loosely consolidated sediments, and solid rocks between a sharply defined upper boundary called the water table and an indefinite lower boundary ranging from a small depth to many thousand feet below the surface. Lower than this even the most persistent fissures must be closed and dry.

Perched Water Tables and Impervious Beds Lying Below the Water Table. — The conditions outlined above are not always completely

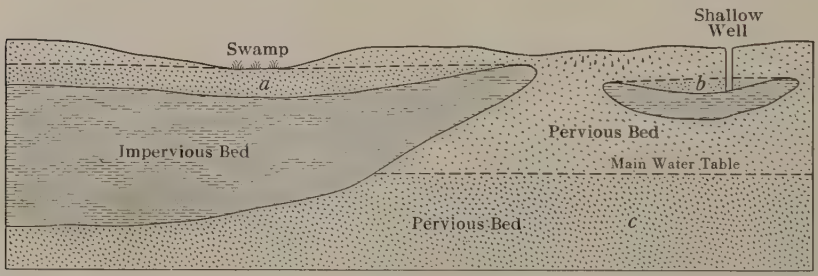

Fig. 100. — Showing perched water tables caused by impervious beds interstratified with pervious beds, and the depression of the main water table by an impervious bed that extends down into the main zone of saturation; *a* and *b* are perched bodies of ground water; *c* is the main zone of saturation. (Adapted from Gregory.)

fulfilled. In regions where the main water table lies at a considerable depth below the surface an impervious stratum in the zone of aëration may hold a local zone of saturation suspended far above the main one. This zone has its own water table called a *perched water table* (Fig. 100). There are also many impervious strata that extend far below the general level of the zone of saturation. A well that penetrated these strata would be dry even after it had passed some distance below the general level of the water table in the surrounding region. Many such impervious strata would cause the water table to be ill-defined.

Return of Ground Water to the Surface

Motions of the Subsurface Water. — Above the water table the movement of subsurface water is directly downward (Fig. 99). Below that level the ground water proper moves in the same general direction as the surface run-off but more slowly, because the friction caused by its passage through the interstices of the rock retards its motion. The continuous addition of parts of this slowly moving sea of ground water to streams and lakes and its seepage at the surface or concentration into springs is the cause of the continuous flow of streams. Were it not for the restraining influence of friction the subsurface flow would be as rapid as that at the surface and every rain would mean a flood and every stream would run dry between storms.

The more deeply the ground water penetrates beneath the surface in uniform material the more resistance it meets due to friction and the more slowly it moves until, at the greatest depths to which it can go, it is stagnant, held motionless by the molecular force of the rocks.

Hillside Springs. — Wherever the surface of the land intercepts the water table, ground water emerges in some fashion. When the

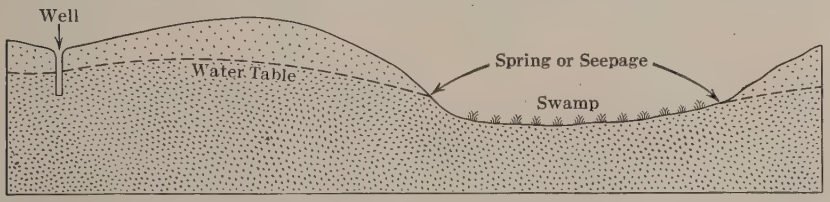

Fig. 101. — Showing conditions favorable for wells, hillside springs, and a swamp or lake.

water oozes out along the line of contact a seepage results, but if the circumstances are such that water is concentrated in quantity sufficient to form a distinct current, an ordinary spring is formed. This is illustrated in Fig. 101. Figure 102 shows the wall of a canyon that has cut the water table so that the water issues in a series of springs.

Wells. — An ordinary well is an opening dug or drilled into the ground to a depth sufficient to penetrate the saturated zone (Fig. 101). The water percolates into the opening from the saturated material around it but does not rise in the well above the level of the water table.

Fissure Springs. — When surface water enters and saturates a pervious, inclined layer confined between two impervious layers the weight of the water in the aquifer generates hydrostatic pressure. If the aquifer is cut by a fissure that leads to the surface, the water will be driven up along the fissure and issue as a spring or a series of springs lined up along

the fissure. Such springs are often called fissure springs. Figure 103 illustrates the following necessary conditions: a catchment area where water may enter the aquifer; a pervious stratum to serve as aquifer

Fig. 102. — Thousand Springs, Snake River canyon, Idaho. The wall of the canyon cuts the water table and allows the ground water to issue as a series of springs. (U. S. Geol. Surv.)

lying between two impervious strata; an inclination of the strata sufficient to cause the water in the pore space to exert pressure on that down slope from it; a fissure to cut off the normal course of the water and furnish a channel to the surface; a point at which the water may issue

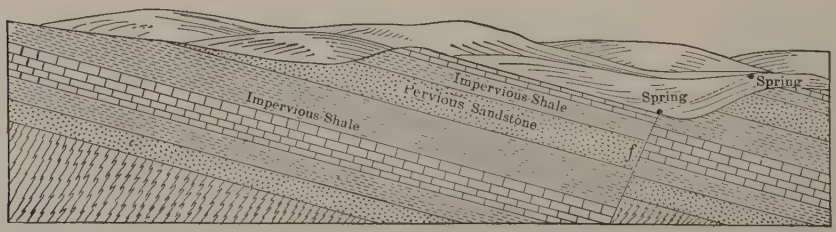

Fig. 103. — Illustrating conditions favorable for springs, if fissures, such as *f*, are present.

below the level at which it entered the aquifer. The difference in elevation between these two levels is known as the *hydraulic head* which, together with the size of the pore space, determines the force with which the water will reach the surface. There is always a considerable loss of head due to friction.

In such an arrangement as is postulated in Fig. 103, springs might be expected at various points along the line of intersection of the fissure with the surface, wherever suitable channels for the upward flow of the water exist. Such springs are usually very steady in their flow and are less affected by droughts than ordinary hillside springs. They are usually cold, but the water may come in contact with heated rocks and issue as warm springs. This is probably the explanation of the warm springs that occur at various places in the Appalachians, as at Hot Springs, Virginia. The water sometimes dissolves unusual amounts of mineral matter and gives rise to *mineral springs*, as at Saratoga, New York; Carlsbad, Bohemia; Bath, England; Wiesbaden, Germany; and Vichy, France. These springs contain considerable though different amounts of sodium, chlorine, carbon dioxide, sulphate, and smaller quantities of calcium, magnesium, and many other elements. There are also springs high in silica, such as Hot Springs, Arkansas; Olette in the eastern Pyrenees; and the geyser waters of Yellowstone Park, Iceland, and New Zealand. It must be realized that, strictly speaking, all springs are mineral springs, since they contain mineral matter in solution. The term is indefinite but is applied in general to those springs that differ markedly from ordinary potable water, either in the quantity of mineral matter in solution or in its character.

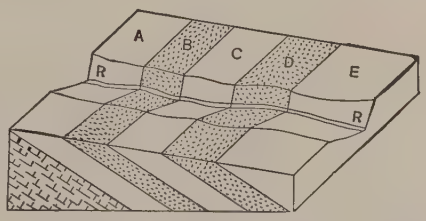

Fig. 104. — To illustrate entrance of water into pervious rock layers, or strata. *ACE*, impervious beds; *BD*, pervious beds; *RR*, course of river.

The condition under which pervious beds become filled with water is important, not only for fissure springs, but also for artesian wells, described below. It is illustrated in Fig. 104. The pervious layers *BD* become filled, not alone by the rain that falls on their exposed surfaces, and by the water that is shed upon them from the higher impervious slopes *A* and *C*, but from the river water that is concentrated from the watershed above.

Artesian Wells. — If under conditions similar to those described above, but in the absence of a fissure, a hole is bored down to the aquifer, the water will rise above the zone of saturation and produce an *artesian well*. The height to which the water rises above the zone of saturation depends upon the pressure, which in turn depends on the height of the water column, or " head " in the aquifer above its upper surface where the well is drilled; the size of the pore space of the aquifer, which determines the loss of head due to friction; and perhaps to some extent

on the weight of the overlying formations which tends to compress the water-bearing stratum and force the water out. In many cases the water rises to the surface or above it. The alternating pervious and impervious beds may have the form of a basin (Fig. 105, A) but that is not a necessary condition. The arrangement illustrated in Fig. 105, B is just as suitable; the water rises through the artificial opening because the diameter of the well is greater than that of the openings that con-

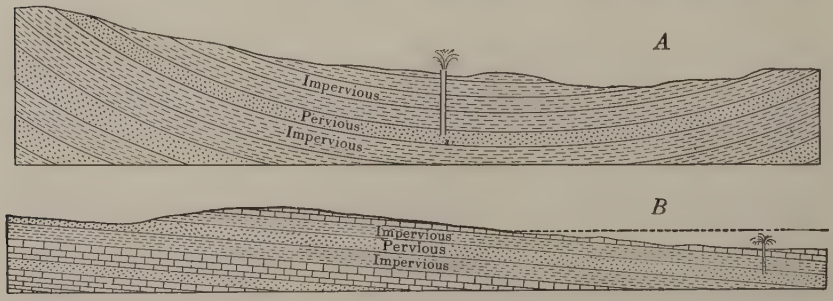

Fig. 105. — Sections showing conditions favorable for artesian wells. Vertical scale exaggerated.

A. Strata are folded so as to form a basin.
B. Strata are inclined in one direction only.

stitute the pore space of the aquifer and the water encounters less resistance than by pursuing its course underground. Figure 106 is a photograph of an artesian well that shows the water spouting up as in a fountain.

The conditions outlined above are possible only in sedimentary strata but the principle is applicable to other rocks as well. It is essential only that the water be confined, that sufficient head be developed, and that fractures or solution cavities act as passages for water. Such conditions are relatively rare and very local. They do not compare in extent with the widespread artesian basins developed in sedimentary strata.

Artesian wells cannot be made simply by boring deeply unless the requisite geologic conditions are present. Deep wells bored into rock so as to intercept the water table are often called artesian wells but this is an incorrect use of the term; there is no difference in principle between wells of this kind and ordinary shallow dug wells. Some of the most important water-bearing formations in the United States which furnish artesian wells are the Dakota sandstone, which comes to the surface along the Rocky Mountains and Black Hills, and underlies large parts of North and South Dakota, Kansas, and Nebraska, and extends into Canada; the Saint Peter sandstone, which outcrops in central Wisconsin, and underlies much of Illinois, Indiana, Iowa, Ohio, Missouri, and Arkan-

sas; and beds of sand that underlie the Atlantic Coastal Plain from Long Island to Texas. The conditions are generally unfavorable for artesian wells in the uplands of New England because the veneer of glacial till and lake and stream deposits is too thin and discontinuous to form artesian aquifers, and it overlies jointed and faulted metamorphic or igneous rocks in which the openings are too closely spaced or the in-

Fig. 106. — Artesian well, near Provo, Utah. (U. S. Geol. Surv.)

dividual fractures are not sufficiently large to establish artesian circulation. In the lowland of central Connecticut numerous fractures prevent any great accumulation of water under pressure in otherwise well-situated sandstone strata. There are exceptions to this rule, however, and small local artesian basins are found scattered throughout the region.

The depth to which wells must be bored before artesian water is attained is very great in some places. In Berlin, St. Louis, and Pittsburgh the necessary depth is about 4000 feet, and depths of 1000 feet are not uncommon. Along the Atlantic coast, on the other hand, artesian wells are generally shallow — from 100 to 300 feet. The volume of water may be very large; the great 12-inch well of St. Augustine, Florida, with a depth of 1400 feet, supplies 10,000,000 gallons a day.

Where many wells are put down close together the drainage basins are likely to interfere with each other, and the withdrawal of water lowers the pressure to such an extent that the water will no longer rise above the level of the aquifer.

GEOLOGIC WORK OF SUBSURFACE WATER

Water underground is an important geologic agent. Its chief work is to *take substances into solution, carry them elsewhere, and perhaps redeposit them;* this work is, therefore, largely chemical in its nature. Although this work may seem insignificant, the total results accomplished during geologic time have been enormous. A part of it has already been considered; thus in the description of the decay of rocks and the formation of soil it was shown that certain constituents such as the alkalies in the rock-forming feldspars go into solution and are removed; and that calcium carbonate, a common rock-making material, is dissolved and carried away by water containing carbon dioxide. These actions are accomplished by atmospheric water as it passes underground, and may thus be regarded as the first stages of the work of subsurface water. Again its work was considered, when it was shown that rivers carry a large part of their burden in solution, and ultimately deliver this material, dissolved by subsurface water, to the sea. Finally the formation of salt lakes, and the deposits that occur in them, illustrate the work of solution, transportation, and deposition.

These facts illustrate the general chemical work of water, partly on the surface and partly underground, but there are certain features that demand particular consideration.

Solution. — The solvent action of rain water passing into the soil and rocks is greatly increased by the substances that it carries with it, or that it may otherwise obtain. In its passage through the air it dissolves carbon dioxide and oxygen, together with minute amounts of other materials, and is thus equipped for doing chemical work (Fig. 107). In passing through the humus and upper soil of humid regions it may absorb much more carbon dioxide as well as organic acids produced by the decomposition of organic matter. In many places, particularly in volcanic regions, volatile substances, especially carbon dioxide, are evolved from the depths, and may dissolve in the subsurface water, thus augmenting the quantity of chemical reagents present in it. In addition, as the water passes into deeper zones the pressure increases and it may come in contact with heated rocks and have its temperature raised, both of which changes greatly increase its chemical efficiency. The amount of gas, such as carbon dioxide, which water can hold in solution, is

directly proportional to the pressure. The amount decreases as the temperature of the water rises but this is more than compensated by the increased chemical activity of hot water.

Pure water dissolves many substances, but with its chemical efficiency heightened as just described, it attacks the minerals composing the rocks and soils with added energy. It takes some of them, such as gypsum ($CaSO_4.2\ H_2O$), directly into solution. In other cases a chemical reaction takes place and new compounds are formed, some of which are soluble and are carried away, while the insoluble material remains.

Fig. 107. — Rock whose more soluble parts are being dissolved by the action of atmospheric waters. Wind aids the rain in removing the loosened material. Near Livingston, Mont. (U. S. Geol. Surv.)

The decay of feldspar, as described under the formation of soil, is a good illustration of this process. The alkaline carbonates produced are leached out, whereas the insoluble clay remains.

The material taken up and held in solution may pursue one or two courses, depending on what happens to the water that contains it. It may work down into the rocks and be deposited there, or it may emerge into the surface drainage and be carried into the ocean.

The process by which the land surface is wasted by solution is known as *chemical denudation*, to distinguish it from the mechanical wear of ordinary erosion. The amount of material so removed each year is very great. Dole and Stabler by assembling a large number of analyses of

the waters of the Mississippi, which analyses give the average percentage of the salts it contains, have calculated that 108 metric tons of matter are removed each year in solution for every square mile that it drains.

Results of Solution of Carbonate Rocks. — Aside from the process of soil formation, in which solution plays an important part, the most obvious results of solvent action are seen in the effects it has upon rocks wholly, or partly, composed of carbonates. The most important rock-forming carbonates are those of calcium, magnesium, and ferrous iron; $CaCO_3$, $MgCO_3$, and $FeCO_3$. The carbonate of calcium especially underlies vast stretches of land, as beds of limestone hundreds or even thousands of feet thick. Besides this, beds of sandstone commonly contain a cement of calcium carbonate that binds the grains of sand together. Since these carbonates, especially calcium carbonate, are attacked by water containing carbon dioxide, many such rock masses must be continually dissolving and wasting away. This is suggested by the fact that in those places where limestone, or calcareous sandstone, is the bedrock the water is always *hard*, i.e., contains lime in solution. Figure 107 illustrates the pitted or cavernous surface of a limestone exposed to the solvent action of water. The soluble material is carried off by the water and the insoluble residue breaks down and is blown away by the wind.

Sinks and Caverns. — In regions where limestone forms the bedrock the surface water works down through joints and fissures and enlarges them by solution. When the water reaches an insoluble layer, such as one of clay or shale, it is stopped in its descent and spreads laterally, finding its way through the rock fissures along the natural drainage slope. These fissures are also enlarged by solution until they become distinct water channels. As the latter enlarge they form *caverns*, while the holes or pipes, leading down to them from the surface above, are termed *sinks* (Fig. 108). Sinks are also formed by the collapse of the roof, weakened by solution, into the cavern beneath.

Some of the individual chambers hollowed out in the rock are 100 feet or more high, and several hundred feet broad. They are connected by intricate passages. The floor on the insoluble stratum may be quite level for long distances though it is broken through in many places when the water excavates new passages and chambers at a lower level (Fig. 109). The final result may be several sets of such rooms and galleries, one above the other. Some well known natural bridges have resulted from partial collapse of cavern roofs.

The limestone regions of the Middle West and the South are noted for their caverns, some of the best known being Mammoth Cave in Kentucky, 10 miles or more long, with 30 miles of winding passages; Wyan-

dotte Cave in Indiana, Luray Caverns in Virginia, and the Carlsbad Cavern, New Mexico. In some places the rocks are honeycombed with passages and almost the entire drainage may pass underground. Large

Fig. 108. — Opening to sink in limestone beds; near Cambria, Wyo. (U. S. Geol. Surv.)

rivers disappear from sight, and after a devious journey below come to the surface again in a different drainage area. They may give rise to huge springs, such as Silver Spring in Florida which has a flow so large

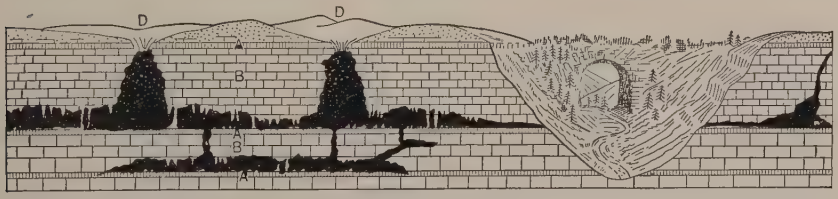

Fig. 109. — Illustrating the formation of caverns and sinks in limestones. *AA*, clay beds; *BB*, limestones. The arch is the remnant of the roof of a former cave, forming a natural bridge. *DD*, sinks, leading to caverns below. (Modified from Shaler.)

that the resulting stream is navigable for small steamers. Other underground streams may be forced up as great springs in the sea, not far from land.

Deposition and Cementation. — From what has been previously stated it is clear that there is an upper belt in the Earth's crust where mechanical and chemical changes and destruction are going on. It is known as the *zone of weathering*, and extends downward to the level of

the water table. Material is being constantly leached out of this zone and carried down in solution into the ground water. This matter is either carried away by the drainage, or deposited in the pores and other cavities in the rocks. The lower limit to which this can proceed is uncertain. It appears to depend on several conditions and it probably varies in different places. Thus the intervention of impervious rock layers, or the charging of the rock pores with gas under pressure, would hinder, or perhaps prevent, further downward movement in a given area. It is in this belt, whatever its thickness, that the sediments are being solidified and cemented by the silica, calcium carbonate (calcite), and other substances deposited in them, and it is, therefore, known as the *zone of cementation.*

There are several reasons why much of the material dissolved by the ground water is redeposited by the same agency within the pores of the rocks. The most important of these are (1) Evaporation of water in open spaces such as caves. (2) Loss of dissolved gas such as CO_2 by warming or evaporation. (3) Cooling of the water. (4) Lowering of the pressure. (5) Chemical reaction between the solution and the rock through which it is percolating.

Replacement. — The deposition of material from solution in ground water does not always take place in openings already present. Frequently the water will dissolve a certain substance and leave behind an equal volume of another substance that it held in solution. Thus a buried tree trunk may be slowly converted into stone by the solution of its vegetable matter and the deposition of opaline silica ($SiO_2.n\ H_2O$). The fact that such replacements take place volume for volume is shown by the complete preservation of the fine texture of the original material regardless of the relative size of the chemical molecules involved.

Concretions. — The capacity of ground water to dissolve and redeposit is nowhere better illustrated than by the formation of concretions. Concretions are regular, rounded, or strangely shaped nodules that occur in sedimentary strata. They are due to deposition from the ground water which has dissolved the substance from other parts of the same rock or brought in some foreign matter. They are explained in more detail in the chapter on sedimentary rocks.

Stylolites. — Irregular, sutured partings characterized by interlocking columns or teeth from a fraction of an inch to several inches in length commonly penetrate soluble rocks such as limestone, dolomite, or marble, essentially parallel to the bedding planes. These partings are known as *stylolites* (Fig. 110). They are formed by unequal solution along the opposite surfaces of a fracture or a clay seam. They are localized by original irregularities of that surface which allow the weight

of the overlying rock to press more at one point than at another and by the varying solubility of different parts of the rock adjacent to the fracture or seam. Once started, the rock opposite the end of each column is dissolved away more rapidly than elsewhere and the teeth penetrate further into the opposite bed and develop striae along their sides. The insoluble material finally collects as a clay cap at the end of each column.

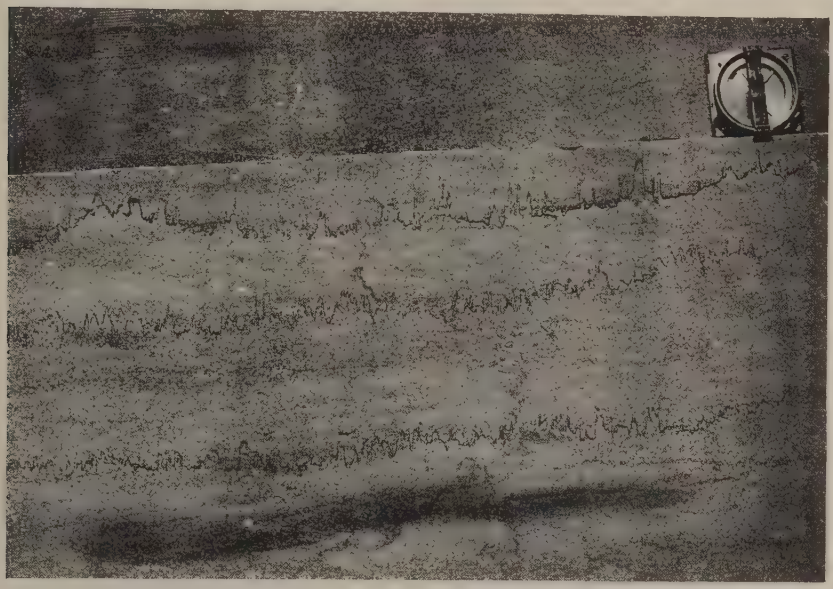

Fig. 110. — Stylolites in Tennessee marble. (U. S. Geol. Surv.)

Some stylolites develop in relatively insoluble rock, such as quartzite. Wherever they occur they indicate the removal of a considerable quantity of rock in solution.

Relation of Subsurface Water to Ore Deposits. — Certain deposits of lead and zinc ores seem to be due directly to the concentrating action of subsurface water that dissolves mineral matter over large areas and concentrates it in a relatively small zone. Many deposits of copper, zinc, gold, and silver are believed to be formed by heated waters ascending from the cooling magmas below — a portion of the total ground water whose quantity we have no means of estimating. Still other deposits — mainly iron ores — owe their value to the removal of worthless mineral matter in solution and the consequent increase in the percentage of the iron minerals in a given area. The value of many important silver and copper veins is due to the concentration of the ore minerals from a large to a relatively small part of the vein by ground

water percolating downward, thus raising the metal content within the narrow zone to a profitable amount.

Deposits in Caves. — The same process that forms caverns also tends to fill them up. For, after they have been opened by solution, the surface waters seeping down through the rock layers which form the roofs dissolve more calcium carbonate, and deposit it in the caverns, producing stalactites, stalagmites, and columns. The manner of their

Fig. 111. — Stalactites, passing below into stalagmites, along a roof-crack. Marengo Cave, Indiana. (U. S. Nat. Mus.)

formation is as follows: a drop of water, charged with calcium carbonate, leaking through to the roof hangs there for a time. While at rest it evaporates a little, loses some carbon dioxide, and consequently deposits some calcium carbonate. Finally, as more water is added from above, it is forced to drop, and falling on the floor below it evaporates still more and leaves another deposit. Thus long, pendant, icicle-like incrustations called *stalactites* grow downward, and broader, dome-shaped masses called *stalagmites* rise up from the floor immediately below. Finally these may increase so that they unite and produce columns. These forms are most likely to develop along fissures in the roof of the cave (Fig. 111).

In past times caves have served as refuges for primitive men who inhabited them, or as dens for wild animals. Because of this the bones of men and animals, stone implements, and other objects have accumulated in them and been sealed up, in the deposits of calcium carbonate on their floors. Relics of this kind, especially in certain parts of Europe, have revealed much concerning the life and degree of culture existing in prehistoric time.

Deposits of Calcium Carbonate by Springs. — The material in solution which is not deposited in the rocks is carried away by the drainage. It sometimes happens that on its way to the sea it comes to the surface

Fig. 112. — One of the terrace formations of the Mammoth Hotsprings, Yellowstone Park, Wyo.

again and is temporarily deposited. Springs that deposit calcium carbonate furnish a good illustration of this. Many springs, especially deep or fissure springs, contain much carbon dioxide gas, under considerable pressure. When this water passes through beds of limestone on its upward journey large quantities of calcium carbonate are taken into solution. On arriving at the surface, partly because of evaporation and partly because of loss of gas through the relief of pressure, the calcium carbonate is deposited and built up into mounds and terraces, some of which are very beautiful. They are illustrated in the basins and terraces of the Mammoth Hotsprings in the Yellowstone Park

(Fig. 112). Other examples of such springs are found in Virginia, Colorado, Banff in Alberta, Carlsbad in Bohemia, Tuscany, and in many other places. Some spring waters contain other mineral substances together with calcium carbonate, or even to its entire exclusion. Many of these springs are used medicinally, as at Saratoga and other health resorts.

In some springs that come from great depth the issuing water is warm, or even hot. This is likely to be the case when the springs occur in regions of active or recently extinct volcanoes such as that in which the Mammoth Hotsprings are situated. In such warm waters the deposit of calcium carbonate may be much increased by the action of primitive forms of vegetable life, the algae, which secrete this substance from the water. In many places the deposition takes place so rapidly that articles suspended in the water become covered in a few days with a coating of the mineral. It is probable that the warmth and chemical activity of the waters of some springs, particularly hot springs in volcanic regions, are greatly increased by gases and vapors that rise from molten magma or masses of hot rock below. Since water vapor is believed to form the largest part of these discharged gases the volume of a spring may be increased in this way.

Nature of Calcium Carbonate Deposits; Travertine, Tufa. — The character of the material formed when calcium carbonate is deposited from solution depends on circumstances, especially on the rate of deposition. When it is produced by slow evaporation, as in the case of stalactites in caves, it is a hard, compact, more or less crystalline substance. *Travertine*, from the old Roman name of a town (Tivoli) in Italy where an extensive formation of the substance exists, is a general name for such deposits. "Mexican onyx" or "onyx marble" is a travertine with banded structure brought out by varied tinting from metallic oxides. When calcium carbonate is formed rapidly from springs the travertine may be porous or loose, or it may coat vegetation and be spongy or mosslike. Such less compact varieties are commonly called *calcareous tufa*, or sometimes *calcareous sinter*. Great deposits of this material are found around the shores of dried-up alkaline lakes, such as Pyramid Lake in Nevada where it encrusts the rocks of the enclosing basin.

It should be clearly borne in mind that these deposits are not *original* formations of calcium carbonate, in the sense in which we think of that word in connection with limestone; they represent in large part previously existent calcium carbonate, such as limestone, or chalk, which has gone into solution, been transferred to another place, and redeposited. They exhibit a temporary stoppage of the material on its way to the sea,

for it is the fate of all deposits of carbonates, if exposed to atmospheric agencies, to be dissolved and eventually taken to the ocean.

Other Deposits by Springs: Iron Oxides, Silica, etc. — Substances other than travertine may be deposited when underground waters issue at the surface. One of these is the hydrated oxide of iron, or, under certain circumstances, iron carbonate. This is a matter of importance because some extensive beds of iron ore have been formed in this way. Under certain conditions silica, sulphur, and gypsum may be deposited, but since agencies other than those which have thus far been described are concerned in the process this matter will be discussed later.

Alkaline Deposits. — The soluble substances that are formed by the decay of the rocks in humid regions are quickly washed out of the soil and are carried by the drainage to the sea. In semiarid and desert regions where the rainfall is scanty there is not sufficient water to perform this function and the salts remain in the soil. At times of rainfall they go into solution, and during the subsequent periods of drought, when the water draws to the surface, they are left behind by its evaporation and form the white incrustation on the soil known as *alkali*. This is a common feature in many parts of the western United States.

The common salts that compose alkali are sodium sulphate, sodium chloride, and sodium carbonate, Na_2SO_4, NaCl, and Na_2CO_3. The name alkali is due to the alkaline reaction and taste of the latter. Magnesium sulphate, $MgSO_4$, and calcium sulphate, or gypsum, $CaSO_4.2\,H_2O$, are often present as well. These salts are not always furnished directly by rock decay; they may have been originally present in beds of sediments laid down in the sea. Their concentration in such arid regions, with inland drainages, gives rise to salt and alkaline lakes. The irrigation of alkali lands, especially if water is too freely or carelessly used, may bring the salts to the surface in such quantities as to injure, or even ruin the land for agriculture.

Mechanical Work of Water Underground; Landslides. — Subsurface water is of little importance as a mechanical agent. It is conceivable that streams running in subterranean channels may erode and transport, but the circumstances that would permit this are exceptional. A more important mechanical function of subsurface water is its aid in causing landslides, both by helping to overcome the friction of masses of rock, soil, and débris lying on steep slopes, and by adding weight to the mass. The saturated masses of soil and rock act like a semifluid substance and, once started from their insecure foundations at times of heavy rainfall or when loosened by earthquake shock, rush down into the valleys causing great damage and considerable changes in the topography. In

high mountains such landslides may precipitate huge trains of broken rock, or talus, down the valley sides, giving rise to rock streams.

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CHAPTER VII

LAKES AND SWAMPS

Inland bodies of standing water are called lakes; expanded portions of rivers are sometimes also referred to as lakes. Small water bodies, particularly where shallow and filled with aquatic plants, are known as ponds, but there is no fixed usage as to these terms. From small ponds there exists every gradation up to Lake Superior, the world's largest freshwater lake, and the Caspian Sea, which though salt is the largest inclosed body of water. Thus, though most lakes contain fresh water, many are salty. The great majority of lakes stand above sea level, but some, including all coastal lagoons, are at sea level; a few, like the Salton Sea and the Dead Sea, are even below. Lakes occur in all parts of the world but since a great proportion of them are a direct result of glaciation, there are more lakes in high latitudes and in high altitudes than elsewhere. There would be no lakes if the surface of the lands were everywhere graded, for then drainage to the sea would be perfect. Where drainage is obstructed and the surface is not at grade lakes occur. Since all the streams are at work destroying obstructions and filling depressions in the ceaseless effort to bring the lands to grade, it follows that lakes are necessarily ephemeral and that all must sooner or later disappear.

Origin of Lake Basins

Warping and Faulting. — Lake basins formed by broad downwarping of the rocks have existed at many times during the Earth's history and most of them probably have been large. We can be sure of this because their shore and bottom deposits are preserved long after the waters that stood above them have been drained away.

Great lake-bearing depressions caused by fracturing and downfaulting of the crust are more common than those caused by folding. The 4000-mile chain of valleys and lakes that includes Jordan River, the Dead Sea, the Red Sea, the upper Nile, and the African lakes, such as Tanganyika and Nyassa, was formed by the sinking down of narrow blocks of the crust, between high steep walls. This great "rift valley" contains more than 30 lakes, several of which are notably large and deep. Lake Tanganyika is about 5100 feet deep, and since its surface is only 2500 feet above sea level its bottom is 2600 feet below. The Platten

See of Hungary, 50 miles long, the Warner Lakes of Oregon, and some of the larger lakes of southern Sweden likewise owe their basins to faulting. In certain cases the formation of such structural basins has occurred within human history in connection with earthquakes. In 1811 an earthquake shook the lower Mississippi Valley and caused such changes in the surface that several new lakes came into existence in the Tennessee portion of the flood plain.

Crater Lakes. — One of the most remarkable natural features in North America is Crater Lake in southwestern Oregon, occupying a crater more than 5 miles in diameter in the summit of the Cascade Range. The arrangement of the rocks which form the crater shows that this was once a high volcanic mountain, but that the top collapsed and sank away, leaving a great *caldera* (see p. 253 and Fig. 176). The depression thus formed filled with rain water, forming a lake 2000 feet deep, without tributaries and entirely dependent on rainfall. The water escapes by evaporation and by seepage into the rocks of the crater rim, reappearing in part as springs at lower levels. Crater lakes are found in most volcanic regions. Notable among them are Lago de Bolsena and others in the great volcanic "campagna" surrounding Rome, Lac Pavin in the volcanic plateau of central France, and the "maaren" of the Eifel in northwestern Germany.

Natural Dams across Valleys. — Natural dams may be thrown across valleys in many ways. Chippewa River, which enters the Mississippi about 60 miles below St. Paul, has deposited sufficient débris athwart the course of the larger stream to restrict flow and the Mississippi has in consequence been ponded for more than 20 miles upstream. Landslides occasionally interrupt the flow of streams, converting them into lakes. Several such lakes are known in the region of the Alps, and a new one was formed at the base of the Gros Ventre Range, in western Wyoming, in 1925. Lava flows occurring in regions where streams have already cut narrow valleys are likely to dam the valleys and form lake basins. The lava barriers are eroded with difficulty because made of solid rock. Lac d'Aydat in central France is of this type, as are several of the lakes surrounding the volcanic cones of Mt. Hood and Mt. St. Helens in Oregon and Washington. Again, where valleys are drowned and the force of their stream currents is dissipated by coastal submergence, it is easy for waves and currents to build bars and barriers across their mouths and so convert them into coastal lagoons. Long coalescent chains of these lagoons are found along the Atlantic coast of the United States from Sandy Hook to Florida.

Glacial Lakes. — The majority of existing lakes are the direct result of glaciation. This is well shown by the fact that most of the lakes of

North America are concentrated within the northern half of the continent, precisely within the area which has recently been glaciated. Some of these lakes occupy depressions scoured or plucked out of the bedrock (cirque lakes are among the most common); others, mere ponds, occupy kettle holes, and still others are ponded back of dumps of glacial débris. Most are small, but some large lake basins as well are of glacial origin. The southern end of Lake Michigan owes its outline to a great crescentic ridge of morainal material behind which it was dammed; while many of the large lakes of the Alps (Lucerne and Constance on the north and Maggiore, Lugano, Como, and Garda on the south) are held in by moraines at the mouths of deep glaciated mountain valleys.

Lakes on Flood Plains. — Flood plains of streams, aside from the stream channels themselves, are areas of conspicuously poor drainage. Shallow lakes are numerous in cut-off and abandoned meanders (*oxbow*

Fig. 113. — A temporary lake in a semiarid region. Wyoming. (U. S. Geol. Surv.)

lakes), in abandoned temporary channels within single large meander loops, and in the depressions formed between natural levees and the outer edges of flood plains. Lake Maurepas above New Orleans is a large lake of the latter type.

Limestone Sinks. — Large areas in southwestern Kentucky and in southern Indiana are remarkable in that they are unglaciated uplands dotted with thousands of tiny lakes and ponds. These lakes occupy limestone sinks and are common also in parts of central Florida, Yucatan, Yugo-Slavia, and other limestone areas of the world. Sinks can contain water only (1) if their bottoms are below the water table or (2) if the outlets are clogged with insoluble clay left after solution of the limestone.

Lake Basins caused by the Wind. — Bare rock surfaces in arid regions such as the high plateaus of northern Arizona and Utah not infrequently contain shallow depressions hollowed out by wind-driven sand. These depressions sometimes contain intermittent lakes. The hollows between live dunes also contain water in some cases, the water being prevented from sinking down through the sand by layers of decaying plant matter. Several large lakes of this kind are found among the dunes at the south end of Lake Michigan, and others are dotted about the Sahara.

Inherited Depressions. — The sea floor contains many irregularities and depressions, and therefore an area newly uplifted from beneath the sea often contains such depressions. Most of the Florida lakes which do not occupy sinks are of this sort. One of the largest is Lake Apopka near Kissimmee.

Playa Lakes. — The broad basins between desert mountain ranges are sometimes filled to a depth of a few feet after the sudden and violent

Fig. 114. — Playa lake in a desert valley near Ludwig, Nevada. The lake is rapidly disappearing through evaporation. (Knopf.)

rainfalls characteristic of such regions. Ephemeral or *playa lakes* are thus formed which disappear a few days or at most a few weeks after the rain has occurred.

Relic Lakes. — Imbedded in the sands and clays of the basin of Lake Champlain, geologists find the shells of marine molluscs and the bones of seals and whales. This is doubly remarkable in that the fossil-bearing deposits now stand 440 feet above the sea. The same fossil remains are found also in the St. Lawrence valley as far west as Lake Ontario. From these facts it is evident that the St. Lawrence and Champlain valleys were occupied in comparatively recent times by an

arm of the sea which has since been forced to withdraw because of uplift of the land. Such lakes as Champlain, relics of former seas, are called *relic lakes*. The Caspian Sea is the largest example of this type.

Life History of Lakes

Lakes are constantly coming into being through the agency of these basin-forming processes, and as constantly those in existence are being destroyed. The broad playa lake in a desert basin is the most ephemeral of all: the waters begin to be sucked up by evaporation as soon as they come to rest. But its containing basin is not destroyed; and with the next cloudburst, the lake reappears as before. In moist regions, however, evaporation goes on much more slowly, and at the same time the inflow of water into lakes is greater. Hence in these regions, lakes do not dry up. But here other forces are at work, which although they operate much more slowly, are far more dangerous to the existence of lakes because they destroy the basins which contain them. "Rivers are the mortal enemies of lakes" (Salisbury). The case of Lake Geneva and the river Rhone illustrates this striking statement. The lake of Geneva occupies a deep mountain valley and is at present about 40 miles long. It was originally 7 miles longer, but each year the turbid Rhone, which enters the lake from the east, brings down great quantities of fine sediment from the glaciers at its source and dumps this material in the quiet water, forming a great delta which fills the upper end of the lake basin from side to side. The water is nearly a thousand feet deep and the delta is correspondingly thick, but its bank-like front is creeping farther westward each year, and in time it will destroy the lake. While this is going on the water which spills from the surface of the lake at its lower end is cutting down its channel and is thus lowering the lake outlet. The water, as it passes under the many bridges of the city of Geneva, is beautifully blue and clear. All of the sediment from the upper Rhone has been added to the delta or has dropped to the bottom during the slow passage of currents down the lake. The lake has thus acted as a great settling-basin depriving the lower Rhone of tools with which to cut. Hence downcutting of the outlet is retarded, but it cannot be stopped. If these processes continue, the western rim of the lake basin will be lowered, and the water level will correspondingly drop, until the lake will be destroyed by this process if it has not already been filled up by the encroaching delta.

Downcutting of the rim of a large lake basin is more forcibly illustrated by Niagara Falls at the outlet of Lake Erie. The water at the brink of the falls is lowering its channel in a bed of limestone. At the

same time the falls are retreating upstream at a rate which averages about 5 feet a year. When the falls have migrated up the river to a point opposite Buffalo, Lake Erie will have been largely drained. It not infrequently occurs that a lake may acquire a wholly new outlet through being tapped by a stream. An example seen by thousands of people every year is found in Yellowstone Lake near the center of Yellowstone Park. In former times this lake was larger than it is now and was drained at a point on its southwest rim by a stream leading through Snake River to the Pacific. The present Yellowstone River, flowing north, was then a small stream engaged in rapidly lengthening its valley headward on a steep slope. The head of this young valley eventually reached the lake, tapped and partially drained it, and diverted its waters into their present course through the Missouri to the Atlantic.

In the South Park of Colorado, in the district west of Pikes Peak, a stream is flowing through a valley whose sides are made of layers of tightly packed volcanic ash. Examination shows that these layers were deposited in a narrow lake, and that they came from an active volcano close by. The delicate ash, sifting down through the lake waters, carefully protected and perfectly preserved the remains of plants and animals that lived in and around the lake. The lake basin was gradually filled with ash and lava and was converted into a stream valley; and from the exposed layers of ash there have been taken more than 1000 species of insects, 250 species of plants (including numerous trees), fishes, and birds, all representative of the life of the time during which the lake existed. The filled-up lake, known as Lake Florissant, has thus become a storehouse of great scientific value.

It is thus apparent that all lakes will sooner or later be destroyed by (1) sedimentation in their basins or (2) drainage at their outlets, or both. To the former factor must be added the contributions made by wind-blown material, the accumulated bones and shells of lake-dwelling animals, and more important still, the remains of aquatic plants. (See pages 163, 164.)

Climatic Control of Lakes

Lakes in Humid Regions. — A depression is formed in a certain locality. Will it contain a lake and if so what kind of a lake? In general the answer depends on rainfall and hence ultimately on climate. Each of the lakes with which we are familiar in eastern North America and in western Europe has streams or springs flowing into it, and an outlet or spillway determined by the lowest point in the rim of the containing basin. Since these regions are humid the rate of evaporation

is relatively slow, and the constant inflow and outflow not only prevent the lake surface from fluctuating greatly but also keep the water fresh. In such lakes plant life is usually abundant, and the accumulations of decaying vegetable matter greatly help to fill up the lake basins, which are likely gradually to become ponds and swamps before they dry up completely.

Lakes in Arid Regions. — In the arid and semiarid regions of the world, where rainfall is slight and evaporation great, lakes rise and fall seasonally, and many dry up and disappear for months at a time.

Fig. 115. — Alkaline salt lake near Parma, Colo. (U. S. Geol. Surv.)

In fact the playa basins in the deserts which lie between the scattered ranges of Nevada, Utah, Arizona, New Mexico, and Sonora contain water for the most part only after sudden and infrequent rains. Here as well as in the desert of Gobi in central Asia, the great deserts of western Australia, the intermontane basins of the Andes, and many other arid regions, evaporation is so rapid and continuous, the soil so porous, and the ground water table so far below the surface, that the streams which start seaward, fed by the rains and snows of high mountains, disappear long before they reach their destination. Some of them dwindle away, spreading out their sediments in great barren flats

covered with incrustations of salt and alkali which in wet seasons become shallow lakes and salt marshes. Other streams, where the water supply is greater, end in lakes. Wherever evaporation prevents the water from overflowing the rims of these interior basins, the lakes either are salt or are fast turning salt. Every river carries dissolved salt in quantities too small to be detected by taste. Slowly but steadily this salt is added to the total already in the water of the lake. It cannot escape by overflow and it cannot escape by evaporation. So year by year the proportion of salt to water increases until not only can it be tasted but

Fig. 116. — Islands of calcareous tufa in Pyramid Lake, Nevada. (U. S. Geol. Surv.)

the water becomes undrinkable. Here appears the great menace of some deserts. The lonely traveler or prospector whose water supply is exhausted, half crazed by heat and thirst, reaches the lake he has seen in the distance. And if the lake is really a lake and not a mirage, he tastes and finds what is worse than no water — water all but saturated with salt. The Great Salt Lake in Utah has a salinity of 18 per cent compared with less than $3\frac{1}{2}$ per cent for the ocean; its waters are so dense that bathers at the beach near Salt Lake City cannot sink, but float buoyantly upon the surface. As a swimmer wades back to the beach the water evaporates from his body, leaving it encrusted with tiny crystals of glistening salt. In the same manner slow shrinkage of the lake is leaving great dry beds of white salt to mark its former gently

sloping shores. The chief salts are common salt (sodium chloride) and sodium sulphate. Calcium carbonate is also present and is commonly deposited as granules on the bottom and shores. In some lakes

Fig. 117. — Salt deposits on the floor of Death Valley, Cal. The salt in the foreground is etched into pinnacles by weathering processes, and is covered with wind-blown dust. Salt formed recently is white. The valley floor is here 250 feet below sea level. The car in the middle distance gives the scale. (Longwell.)

this substance is deposited in great spongy, mosslike masses known as *calcareous tufa*. By covering large rocks with thick incrustations, this material produces many curious and striking forms (Fig. 116).

Fig. 118. — Former shore lines and wave-cut terraces of the ancient Lake Bonneville.

Extinct Lakes. — Observant travelers in Utah and Nevada have had their curiosity aroused by frequent sights of parallel rows of cliffs and great flat-topped terraces forming huge flights of steps up the moun-

tain sides to levels as high as 1000 feet above the valley floors. When examined in detail, these " steps " prove to be wave-cut cliffs and wave-built terraces, bars, and beaches, as well as deltas high and dry. These features are arranged in sets at different levels, and when traced along the various mountain ranges of the region, the highest set is found to mark the shoreline of a lake which must have had an area of 20,000 square miles (nearly as great as that of Lake Michigan) and a maximum depth of more than 1000 feet. The water must have been 850 feet deep at one time above the site of the Mormon Temple in Salt Lake City. The outlet was to the north, through Red Rock Pass to the Snake River, and thence through the Columbia to the Pacific. Since there are several sets of shorelines at different levels, we must conclude that the lake dwindled by stages from its former gigantic size to that of the Great Salt Lake, which represents the deepest pool in the bottom of the basin of its predecessor. Similarly, Pyramid Lake and its neighbors in Nevada are the residual pools of another huge water body, which rivaled in size the present Lake Erie. The Utah lake has been named *Bonneville* and the Nevada lake *Lahontan.* The fact that some of the terraces of Lake Bonneville are associated with glacial deposits indicates that the lake had its origin in the Glacial Period. The presence of the ice must have made the climate much more humid than now, because since the ice retreated the outlets have been slowly cut down and at the same time the lakes have been gradually drying up. Salt deposits interbedded with the lake clays, all now high

Fig. 119. — Map of the former Lake Bonneville; ruled areas show present water-bodies.

and dry, bear witness to the evaporation which has been going on during the time since the Glacial Period. Such deposits, so clearly related to the lakes, help us to understand that climates slowly change; and when we find beds of salt in the rocks of central Michigan, a humid region, we can be fairly certain that the climate of that region was dry at the time the salt beds were being deposited (Figs. 118, 119).

The Salton Sea. — The main line of the Southern Pacific Railroad eastward from Los Angeles has the curious distinction of running for more than 60 miles through an area that lies below the level of the sea.

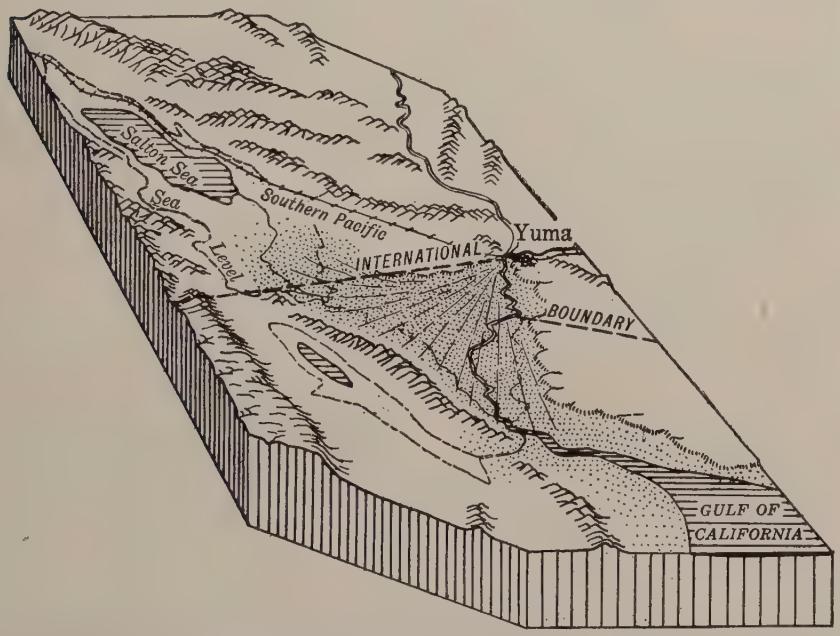

Fig. 120. — Delta of the Colorado River and the Salton Basin. Dashed lines show sea level. Longest dimension of block about 200 miles.

The area is the Salton Basin. It is a basin without outlet, and its bottom is 273 feet below sea level. In its center is a shallow salt lake known as the Salton Sea (Fig. 120). In the recent geologic past the floor of this basin sank, and would have filled with sea water; but Colorado River, heavy with silt, discharged into the Gulf near what is now the southern end of the Imperial Valley and built up a delta and fan so large that they formed an effective dam at the head of the Gulf. Like other deltas this one was trenched by shifting distributary channels of the river, which discharged water sometimes into the shortened Gulf and occasionally into the isolated basin, filling it with water.

When dry, the basin was kept virtually a desert under the prevailing arid climate. The last natural discharge into the basin occurred perhaps 300 to 1000 years ago. In 1900 the trapped water had dwindled to a small salt lake. At this time the fertility of the alluvial soil was realized, the basin area was settled, and an irrigation canal was dug from Colorado River to the bottom of the basin. Since the floor of the basin was well below the baselevel of the river, the gradient of the canal was greater than the gradient of the river between its mouth and the head of the canal. The flow of water into the canal was inadequately controlled by headworks designed to be protective. In 1905 the Colorado rose in flood, overtopped the headworks, poured into the canal, and following the new steep gradient, cut a great trench, in some places 80 feet deep, swelled the Salton Sea to many times its former size and depth, and flooded the railroad right of way for more than 40 miles. The inundation of the basin resisted all efforts at permanent control for more than 18 months; but it was finally mastered. The railroad tracks were shifted from 200 feet below sea level to 150 feet below sea level as a precaution against possible later floods. The Salton Sea began to dwindle as soon as the abnormal supply of water was cut off, but now the waste water from irrigation keeps the level nearly constant.

Indirect Functions of Lakes

Large lakes are effective in modifying local climates by increasing atmospheric humidity and by cooling the air in summer and warming it in winter. Lakes of all sizes are very important as regulators of stream flow, acting as storage reservoirs and minimizing the height of floods in lower regions to which they are tributary. This fact has been recognized by the Egyptian Government in its project to increase artificially the size of Lake Tana in Abyssinia, one of the important sources of the Nile, in order to increase its effectiveness as a regulator of water supply in the Lower Nile Valley, an area of great economic importance. Similarly, suggestions have been made in the United States that a series of artificial lakes be constructed in the Mississippi drainage basin in the attempt to control near their sources such floods as caused the disaster of 1927. All lakes likewise act as settling basins for river sediment. Most of the streams tributary to Lake Erie are well loaded with mud, but the clarity of the water which pours out of the lake and over Niagara Falls bears testimony to the amount of material which is constantly being dropped upon the lake bottom.

SWAMPS

Swamps are areas of saturated ground. The majority of swamps represent a stage intermediate between lakes or ponds and dry land. Most lakes will in time become swamps, and many shallow basins alternately contain swamps and lakes according to the season. Swamps are commonly found in three types of regions, but these regions by no means exhaust the possibilities. (1) Swamps are both numerous and large on nearly level coastal plains which are former sea floors slightly uplifted.

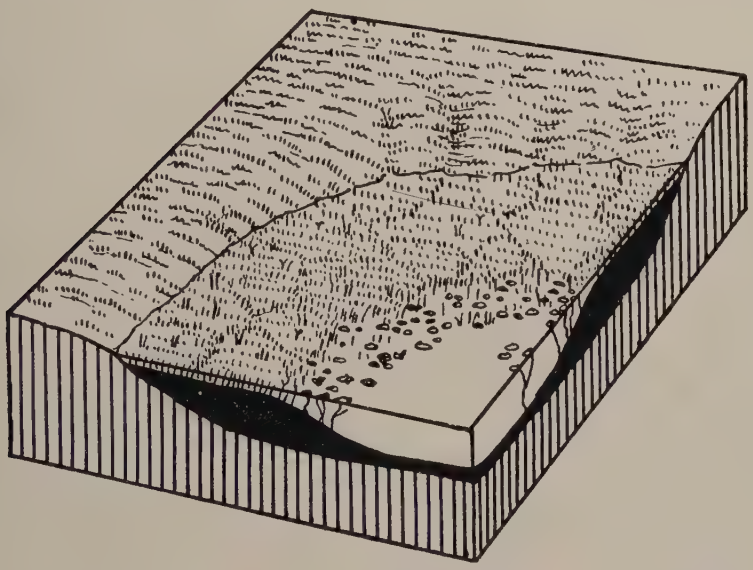

Fig. 121. — Destruction of a small lake by formation of peat. The accumulating layer of peat (solid black) is fringed by aquatic vegetation consisting of water weeds and pond lilies while directly above it are encroaching semi-aquatic plants, mosses, and bushes.

Such swamps are distinguished from salt marshes and are almost continuous along the South Atlantic and Gulf coasts of the United States, chief among them being the Dismal Swamp in Virginia and North Carolina, and the Everglades in southern Florida. Some of these swamps may be old lagoons, uplifted together with the offshore bars by which they were shut off from the sea. (2) Flood plains and deltas with their basins formed by old channels and by natural levees contain great areas of swamp land. (3) Broad glaciated areas such as the greater part of the Great Lakes region of the United States, northeastern Canada, and the Baltic Plain of northern Germany are dotted with swamps, most of them small.

Formation of Peat. In humid regions, the shores of small lakes and protected coves of large lakes support an abundance of aquatic vegetation such as pond lilies, water weeds, and rushes. As these plants die, their substance decays in the water by slow oxidation, which is caused largely by the activities of bacteria. During the metabolism of the bacteria, waste products, antiseptic in their action,

Fig. 122. — Dismal Swamp, Va. The projections from the cypress roots serve to give them air; they extend downward into the mud and help to anchor the tree in the semi-liquid mass of the bog. (U. S. Geol. Surv.)

are excreted; hence when this waste matter reaches a certain concentration in the lake water the bacteria can no longer exist, further decay is prevented, and the partially decomposed matter is preserved. This matter is brownish or blackish, has a high carbon content, and is known as *peat*.

As peat is gradually built up along the shore of a lake, newer generations of aquatic plants advance toward the center of the lake, and other types of vegetation such as mosses encroach over the peaty area which was formerly water. Thus the lake, surrounded by concentric belts of different kinds of plants, gradually decreases in size until it is obliterated, and a bog floored with a thick accumulation of peat takes its place (Fig. 121).

Economic Aspects of Peat. — In many countries, especially in Europe, peat is cut from the bogs, dried, and used as a domestic fuel. In America peat has hitherto been little used because of an abundance of coal and wood. The peat resources in bog lands within the United States, however, are enormous, and will probably constitute a valuable source of fuel in the future.

Peat is of special interest to the geologist in that it represents the first stage in the transformation of vegetable matter into coal. All gradations may be observed from peat through lignite, bituminous (" soft ") coal, and anthracite (" hard ") coal. Thus if present peat bogs, the best of which lie in temperate and cold humid regions, were left untouched, many of them in the course of time would be covered with sediment, the remaining necessary changes would take place, and the result would be the formation of coal interbedded with the rocks of the Earth's crust. So it appears that as we use the coal formed many millions of years ago, potential coal is now forming for use many millions of years hence.

Quaking Bogs. — In the lakes and swamps of cool temperate and cold climates there flourishes a plant known as *sphagnum moss*. Sphagnum grows abundantly in the northern United States, Canada, and northern Europe, giving these northern swamps an aspect different from the swamps of the south. It readily grows outward on the water surface of a small lake, forming a floating mat which conceals clear water and black liquid muck beneath. When one walks on the mossy surface, the whole mass shakes and quivers; hence these bogs are called *quaking bogs*. Men and animals have not infrequently fallen through such unstable surface mats and have been lost in the quagmires below. Since the antiseptic nature of the bogs, deadly to bacteria, largely prevents the decomposition of organic matter, the bodies of animals entombed many thousands of years ago are dug up in remarkable states of preservation. In the state of New York alone the remains of more than 200 elephants which became mired during or after the Great Ice Age have been dug up from peat bogs.

Economic Value of Swamp Lands. — Peat is not the only valuable deposit of swamp origin. In lakes and swamps as well as in warm springs and even in the surface waters of the ocean live great swarms of microscopic plants called *diatoms*, which secrete minute shells of silica. When these organisms die, their tiny shells fall to the bottom and build up a white, porous deposit known as *diatomaceous earth*. Beds of this material, some of them hundreds of feet in thickness, are now excavated and used in the manufacture of dynamite, polishing powder, and other products.

Certain valuable deposits of iron ore can be traced back to their origin in lakes and swamps. Many areas of marsh and standing water today are inhabited by countless numbers of microscopic organisms known as *iron bacteria*, living wherever the water contains dissolved iron. As an essential part of their life process, they secrete the iron from solution, convert it into insoluble form, and precipitate it, causing an accumulation of iron (usually in the form of limonite) on the lake or swamp floor. The bacteria, though minute, exist in such inconceivable numbers, and in successive generations work through such long periods of time, that large deposits of iron are gradually built up. These de-

Fig. 123. — Reclaimed land from the Dismal Swamp, Va. (U. S. Geol. Surv.)

posits where commercially important are known as one type of "bog iron ores." They are particularly abundant in the glaciated regions of Europe, Asia, and North America, although they are now of very slight importance in the mining industry.

The agricultural value of swamp land when drained is not to be minimized since swamp soils are highly fertile. The swamps and bog lands in the United States have a combined area greater than the area of New England, and with proper draining much of this land could be reclaimed for farming. Drainage canals and ditches have been used for centuries in Flanders and Holland; and now in many parts of the United States large areas of wet land are being prepared in this way for agricultural use. Thus artificial means are being used to accomplish what in time would have been achieved by normal erosion — more complete gradation of the land.

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CHAPTER VIII

OCEANS AND SEAS

General Features

Geologic Rôle of the Oceans and Seas. — Marine waters cover about three-quarters of the globe. Their influence on the geologic history of the Earth's surface has been both direct and indirect. The importance of the oceans in regulating climate, for example, is well known. It is a remarkable fact that throughout the hundreds of millions of years of geologic history, the general temperature of the Earth has never fallen below the freezing point nor exceeded the boiling point of water. For this remarkable constancy we must thank the unflagging energy of the sun's warmth, but the oceans have also undoubtedly served as a great stabilizer by storing up excess warmth against times of lessened solar radiation. Ocean currents carrying warm waters into the higher latitudes and returning the colder waters toward the equator also aid greatly in distributing the heat and softening the contrasts between climatic zones. Moreover, evaporation from the ocean's surface in the last analysis supplies all of the moisture borne by the winds to fall as rain and snow upon the lands. The intense aridity of the basins of the several continents, as the great desert of Central Asia, the deserts of Australia, the Sahara, and the Great Basin of the western United States, suggest what might be expected if the Earth had limited oceans.

Through its waves and currents the marine water is an energetic agent of erosion, gnawing away relentlessly at the margins of all the lands. Moreover, the seas and the oceans are the final settling reservoir in which are deposited the sediments brought down by the rivers, as well as those produced by marine erosion. Upon the sea floor these sediments are shifted about to fill the depressions, or, in the shallow places, tossed up to build out the coast lines of the lands and ultimately to be compacted and cemented into sedimentary rocks.

Relations of Oceans and Continents. — If the waters were withdrawn from the ocean basins, we should see that the grandest relief features of Earth's rocky crust are not its mountain ranges but the uplifted continental masses that lie as vast plateaus 3 miles above the enormous plains of the ocean floor. The naked face of the moon presents to us a

spectacle of this sort, for its maria or " seas " would be such in reality if there were water upon its surface.

Presumably the continental masses stand high because they are made of lighter, granitic rocks and the oceanic areas are depressed because they are formed of heavier, basaltic rocks. On the moon, the depressed segments are relatively small; but Earth's oceans are far greater than its continental areas. It is apparently only a coincidence that there is just sufficient water on the Earth to fill the ocean basins brim full and place the shoreline upon the margin of the continents.

Size of the Oceans. — Not only do the ocean basins have three times the area of the continents, but they attain a maximum depth of more than 6 miles (35,410 feet) and an average of about $2\frac{1}{2}$ miles (13,000 feet), whereas the average elevation of the emergent continents is only about half a mile above sea level. So vast is the volume of oceanic waters that if the continents were planed down and leveled into the basins, a universal ocean would cover the entire Earth to a depth of more than $1\frac{1}{2}$ miles. It is important to realize, however, that even this depth is slight in comparison with the diameter of the Earth. If, for example, a globe 3 feet in diameter were dipped into water and then withdrawn, the film of wetness adhering to it would represent to true scale an ocean half a mile in depth, and if, in drying, the globe should warp by so small an amount as to lessen its diameter at any place by one-hundredth of an inch, the change would correspond to the depression of one of the major ocean basins. It is evident, therefore, that relatively trivial warping of the Earth's crust as a whole would suffice to deepen parts of the ocean basins and draw off the water from the continents or, by elevating parts of the ocean floor, to cause an overflow of the lower land areas, with vast marine inundations. Whatever the cause, such changes have occurred many times in the geologic past. Some of them will be described in Part II of this book.

At present the oceans more than fill their basins and flood the margins of all the continents. These slightly submerged borders of the continental masses are known as continental shelves (Fig. 124). They are widened on the one hand through landward planation by the sea and on the other through the deposition of sediments swept toward the ocean by marine currents. They are broad along stable coasts where these processes have operated without interruption for long geologic ages, but commonly are narrow where young mountains have been formed near the continental margin. Thus, for example, the continental shelf is 60 to 80 miles wide on the Atlantic coast off the Carolinas and only 10 to 25 miles, or even less, along the Pacific coast of California.

Although it is customary to use sea level as a common datum of

reference, it should be noted that the surface of the oceans is not a perfect sphere. Its polar diameter is about 27 miles less than the equatorial and there are, in addition, local and irregular departures from the spherical surface due to the fact that the gravitational attraction of the continental masses draws the water up about the shores somewhat as surface tension draws up the margins of the water in a vessel. Thus the sea level is higher on the coasts than far out in the ocean and not as high on low coasts as on those bordered by high land masses, like

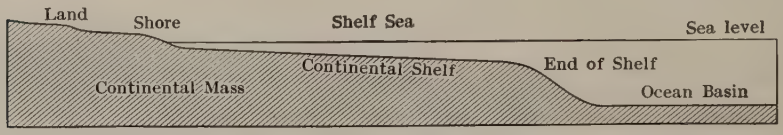

Fig. 124. — Section through edge of continent into ocean basin. Vertical scale greatly exaggerated, causing slopes to appear much steeper than they actually are.

the Andes. It is not improbable that such distortion of the sea level amounts, at its maximum, to several hundred feet.

Oceans Versus Seas. — In common usage the words sea and ocean are synonymous, and we speak of deep-sea deposits, and of going to sea, when actually we have in mind the ocean. There is, however, a technical distinction observed by scientists who apply the term *ocean* only to those five vast bodies of deep water that occupy the ocean basins proper and lie *between* the continental masses; namely, the Atlantic, the Pacific, the Indian, the Arctic, and the Antarctic. The seas, on the contrary, are relatively shallow bodies of marine water that lie *upon* the continental platforms as, for example, the North Sea or Hudson Bay. This technical restriction of the term sea accords with its original use by the peoples of northwestern Europe, who applied it to the North Sea and the Baltic in contradistinction to the Atlantic or outer ocean.

The distinction is one of great significance because of the profound influence of depth on the processes that operate on the ocean floors. The seas are for the most part less than 600 feet deep, whereas the average depth of the oceans is more than twenty times as great. The floor of the seas is a region of activity, touched by the warmth of sunlight and stirred by waves and currents that shift its sediments and bring oxygen and land-derived food to its teeming life. It has been the cradle of evolution for the lower forms of life and the scene of geologic processes of profound importance. The great oceanic floor, on the contrary, is a realm of dreary monotony. Except where submarine volcanoes or earthquake lines are active, nothing disturbs its quiet save the noiseless descent of wind-blown or meteoric dust or of derelicts of the surface

life miles above; and its denizens eke out an existence in utter darkness and cold.

The Mediterranean "Sea" fits the definition of neither sea nor ocean, for it is of oceanic depth and yet is comparable in size to a sea and closely circumscribed by lands. The term *mediterranean* is therefore used in a generic sense to include such circumscribed bodies of deep water lying between continents. The Caribbean "Sea" is thus another mediterranean.

Classes of Seas. — The seas may be arranged, for convenience of reference, in three classes: marginal, epeiric, and relic.

Marginal seas are those which lie upon the continental shelf and are more or less openly connected with the ocean. Where they are somewhat delimited by projecting lands, various portions of the marginal seas bear distinct names; as the North Sea, the Yellow Sea, the Gulf of St. Lawrence, Cape Cod Bay, or the Gulf of Maine. The shallow water overlying the continental shelf of the central and southern Atlantic coast of the United States is just as much a sea, but it and others similar have not been named.

The marginal seas fall naturally into two further subdivisions. Those lying upon the continental shelf proper are known as *shelf seas*. These, including the examples cited above, are shallow. The other group, known as *funnel seas* (Grabau), occupy strong structural depressions that trespass upon the continental border, like the Gulf of Lower California, the Bay of Bengal, and the Arabian Sea. They are far deeper than the typical shelf seas, especially at the outer ends where they are less sharply marked off from the ocean basin proper. They partake of the characters of mediterraneans but are less distinctly hemmed in by lands.

The *epeiric seas* [Gr. *epeiros*, a continent] are those that lie so far in upon the continent as to be largely land-locked. At the present time there are but two good examples, the Baltic Sea and Hudson Bay, but during certain past geologic ages, when the lower portions of the continents became widely flooded, epeiric seas were of vast extent and great importance. In fact most of the sedimentary rocks upon the present lands were formed in epeiric seas.

The *relic seas* are those like the Caspian, which, through crustal uplift, have become isolated from the oceans. They are, indeed, great brackish lakes and should be so called; but their former marine connections may be clearly indicated by the life they still harbor. For example, although the waters of the Caspian are now almost fresh, due to the inflow of great rivers, its fauna includes marine fishes, porpoises, and seals. It is known to have been connected with the Arctic Ocean as

well as the Mediterranean within rather recent geologic time. Lake Champlain is also a relic sea, for it was flooded by marine waters from the Gulf of St. Lawrence a few thousand years ago, and fossil marine shells are abundant in places in the clay above its present shores.

Depth Zones. — The depth of the water is one of the most important factors in the marine environment for it conditions the amount of light and warmth so vital to marine life, and also the effectiveness of the waves and currents that determine the character of the sediments. From this point of view, the marine realm may be subdivided into several bathymetric or depth zones (Fig. 125).

The *littoral zone* extends between extremes of high and low tides and includes, besides the actual shore slope, the mud flats, laid bare at low

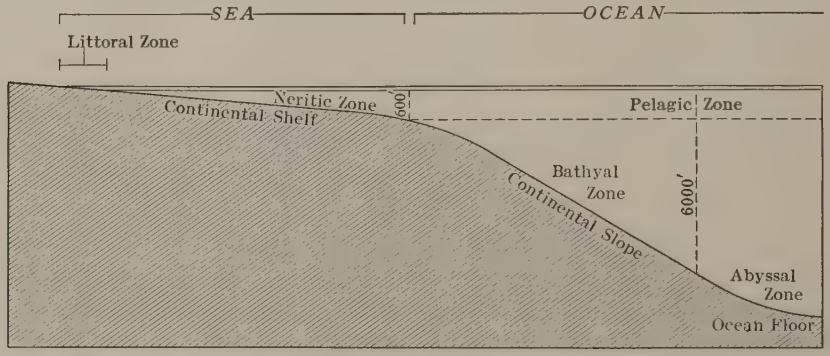

Fig. 125. — To show the several depth zones of the marine realm. The vertical scale and the steepness of slopes are much exaggerated, and the full depth of the ocean is not shown.

tide, and the salt-water marshes, which are flooded only at highest tide. The entire width of the littoral zone is usually not over 2 miles and as a rule it is much less so that its total area is only about 60,000 square miles. In a few regions of exceptionally high tides or of rapid silting, however, the littoral zone reaches a considerable width. For example in the Bay of Fundy, where the average tide rises between 30 and 40 feet, the mud flats at ebb tide are 4 or 5 miles wide. In parts of the Mississippi River delta the salt-water marshes attain a width of fully 25 miles.

The environment of the littoral zone is more varied than that of any other part of the sea floor. Alternately covered by salt water and laid bare to the atmosphere, it may be baked in sunshine one hour, and deluged by the fresh water of a storm the next. Moreover, the waves break here with their greatest vigor, and wave and current action is strongest. At any one place the environment changes with the tides

and storms from day to day, and it is equally variable from place to place along the shoreline, depending on the degree of exposure to the open sea. Accordingly the littoral region is the most difficult life zone, and relatively few kinds of animals live here though some, like the barnacles and crabs, exist in vast numbers.

The *neritic zone* includes the shallow, sublittoral sea floor from the limit of low tide out to the margin of the continental shelf; that is, to a depth of about 600 feet. It is known to the Germans as the Flachsee (flat sea) and, indeed, its floor is in general remarkably even, for it is mantled by the sediments derived from the land and shifted by the waves and currents into the deeper places, smoothing out the inequalities of its surface. The average width of the neritic zone being about 75 miles, its seaward slope averages less than 10 feet to the mile, and it would appear to the eye as an utterly monotonous level plain. Only locally, and generally near the shore, does it reach an inclination as great as 50 feet to the mile.

The total area of the neritic zone is about 10,000,000 square miles, or approximately one-fourth that of the land areas of the world. In past geologic ages, however, the zone has been greatly expanded, as the lower parts of the land areas were extensively flooded by epeiric seas.

The neritic zone is a realm of change and activity. Waves and currents keep the water in motion, salinity and muddiness vary from place to place, and the temperature changes with the seasons and varies with the latitude. Sunlight penetrates to the bottom. It is the most hospitable and stimulating environment for sea life and the teeming host of creatures that live here play an important geologic rôle.

Beyond the edge of the continental shelf there is a somewhat steeper incline, known as the *continental slope*, that is fairly well defined from a depth of 600 to about 6000 feet, where it imperceptibly grades into the abyss of the deep ocean. The intermediate depths that cover the continental slope constitute the *bathyal zone*. Due to the greater depth, bottom currents are feeble here and no coarse sediment can be introduced; but fine, land-derived muds cover a vast expanse of about 18,000,000 square miles, an area exceeding one-third that of all the lands.

The *abyssal zone* includes the whole of the ocean bottom below a depth of 6000 feet. Its average depth is about 13,000 feet and in general it is monotonously flat, lacking the smaller relief features such as the hills and valleys of the land. Nevertheless on a large scale there are swells and depressions rising above and sinking below the general level of the ocean floor. The depressed areas, if more than 18,000 feet below sea level, are known as *deeps* and of these 57 have now been discovered. The deeps are of two rather sharply distinct classes: (1) the vast basin-

like depressions with irregular borders that form the central portions of the several oceans; and (2) narrow and troughlike depressions, mostly situated near the continental margins and several of them parallel to marginal mountain ranges. A good example of the latter is the Tuscarora Deep, paralleling the Japanese Islands, with a depth of about 28,000 feet. These marginal deeps, or *foredeeps*, appear to be areas that have been depressed by breaking or sharp bending of the ocean floor to compensate for the uplifted marginal mountains.

The greatest depth known is that of the Swire Deep, which lies along the eastern side of the Philippine Islands. About 50 miles east by north of Mindanao, a depth of 35,410 feet was discovered by the cruiser "Emden" in 1927 and numerous other soundings in the vicinity show approximately 6 miles of depth. In the Atlantic, which in general is shallower than the Pacific, the Nares Deep, off Porto Rico, holds the known record with 27,972 feet. These great deeps of the ocean floor correspond, in area and in magnitude, to the highest elevations on the land, the oceanic depths attaining more than $6\frac{1}{2}$ miles below sea level and the loftiest mountain range, the Himalaya, about $5\frac{1}{2}$ miles above.

The floor of the deep ocean presents an inhospitable environment for living things, since at this depth no sunlight ever penetrates the utter darkness; the pressure amounts to about 1 ton per square inch for each mile of depth; and the temperature is less than 4° C. Life is sparse and grotesquely specialized. In the absence of sunlight, no plant life exists and the animals are either carnivores or else scavengers, feeding on the dead organisms that settle down from the surface layers of the ocean.

Over the abyssal floor, land-derived sediments are wanting, but everywhere there is a covering of peculiar deposits known as *oozes*. They are so soft and fine that water movements of one-half mile per day are sufficient to shift them on the bottom. These fine sediments and their sources are discussed on page 206.

The surface waters of the open oceans constitute the *pelagic zone* or realm [Gr. *pelagos*, the open sea]. The organisms that live here enjoy the warmth and sunlight but must perpetually swim or float. As indicated above, they may contribute to the bottom deposits of the ocean basins.

Composition of Marine Water. — A barrel of sea water contains about 12 pounds of dissolved mineral salts. These dissolved salts constitute about $3\frac{1}{2}$ per cent of the weight of marine waters. Chemical analyses from all parts of the oceans, and from different depths, show that the composition of the water in the open ocean is remarkably

uniform, and that over 99 per cent of the dissolved mineral matter probably represents only a half dozen common salts, as follows:

Sodium chloride, $NaCl$	77.8 per cent
Magnesium chloride, $MgCl_2$	10.9 per cent
Magnesium sulphate, $MgSO_4$	4.7 per cent
Calcium sulphate, $CaSO_4$	3.6 per cent
Potassium sulphate, K_2SO_4	2.5 per cent
Calcium carbonate, $CaCO_3$	0.3 per cent
Minor constituents	0.2 per cent
	100.0 per cent

Among the minor constituents are traces of a surprising number of the chemical elements, including fluorine, boron, arsenic, nitrogen, phosphorus, silicon, radium, copper, iron, lead, silver, and gold. Most of these occur in such small amounts, however, that they are of no direct geologic importance.

Although common salt ($NaCl$) constitutes more than three-quarters of the dissolved mineral matter, there is little more than a trace of calcium carbonate ($CaCO_3$), the ratio of salt to calcium carbonate being almost 260 to 1. This relation is the more striking when it is recalled that river water, whence the oceanic salts have been derived, normally carries far more calcium carbonate than sodium chloride. The discrepancy is due, of course, to the fact that $CaCO_3$ is either used by organisms or chemically precipitated about as fast as it is delivered to the sea, whereas nearly all the salt ($NaCl$) that has been brought down to the sea since the beginning of time remains in solution. The total quantity of the dissolved salts is astonishing, for it amounts to about 32,000 million million tons, and if precipitated and crystallized into a bed of solid salt, would be sufficient to cover the whole of the United States to a depth of more than $1\frac{1}{2}$ miles.

Silica and nitrogen, though present only in traces in marine waters, are of vast biologic consequence. The nitrogen is required as food and the silica as shell material by the single-celled plants known as diatoms, which float near the surface in incredible numbers, and form, to a large degree, the ultimate food supply for all the animals of the marine realm. Diatoms flourish and multiply rapidly if the supply of dissolved nitrogen and silica is enriched but their expansion is checked by a falling off of these foods, so that the " pastures of the sea " may be said to depend on the supply of nitrogen and silica. The latter is one of the common minerals in river water and its rarity in the sea is probably the result of its use and precipitation by organisms.

In addition to the mineral salts, marine water holds in solution oxygen, nitrogen, carbon dioxide (CO_2), and other gases. As a rough average, 1000 cu. cm. of sea water holds in solution about 20 cu. cm. of air (oxygen and nitrogen) and 23 cu. cm. of carbon dioxide, both measured at standard temperature and pressure. The importance of these gases is out of all proportion to their quantity. All the animal life in the seas and oceans requires oxygen. Oxygen is necessary, moreover, for the oxidation and destruction of the carcasses of dead organisms on the ocean floor, and it thus aids in keeping the water clean and habitable. In places, such as the depths of the Black Sea, where defective vertical circulation fails to provide a sufficient supply of oxygen, the water becomes putrid and foul with the gases of decay, and only anaërobic bacteria are able to live. The entire ocean bottom would be equally forbidding if it were not for the bountiful supply of dissolved oxygen.

Carbon dioxide is the primary food stuff from which green plants draw their sustenance and in turn build up the organic foods upon which animal life is dependent. Without it no life would be possible. At the same time it aids in the solution of the calcium salt, $CaCO_3$, as explained in Chapter III. Since the solubility of CO_2 varies inversely as the temperature, the cold waters of the polar regions and of the deeper ocean bottom can hold more CO_2 than the warmer surface waters. It is this excess of CO_2 in the abyssal depths that causes the solution of all limy shells that settle there.

Movements of Marine Waters

Waves. — The common movements of marine water include the waves, tides, and a variety of currents. Probably the most important and the most direct of these in their geologic effect are the waves.

The ordinary waves of the sea are generated by the wind blowing in irregular gusts and pressing unevenly upon the surface of the water, which is thereby thrown into little undulations. Once formed, these undulations are maintained and increased by the pressure against their windward side and so are driven forward in endless succession.

It is important to realize that the *wave form* travels ahead, independent of the movement of the water itself, just as waves may ripple across a field of standing grain though the individual stalks merely bow as each wave passes and then return to their original positions. Indeed, if the water actually rushed forward with the velocity of storm waves, the oceans would hardly be navigable. The path of movement of an individual particle of water is almost a closed circle (Fig. 126), for it rises and rides forward with the crest of the wave only to slide back into the next

trough to its original position, a fact that can be observed by watching a bit of cork or driftwood as the waves pass under it. As a matter of fact, due to the friction of the wind, it may not return quite to its original position but may be slightly advanced by each wave. In this manner, *wave-formed currents* are generated.

The height (i.e., vertical elevation of crests above trough) and the length (i.e., horizontal distance from crest to crest) of the waves increase with the velocity of the wind, its duration in a given direction, and the length of " fetch " across open water. For this reason small bodies of water and protected embayments of the coast are never affected by great waves. A distance of 400 or 500 miles is sufficient to produce the greatest waves, however, since the wind does not blow steadily in any one direction over greater distances.

Size and Velocity of Storm Waves. — Observations have shown that storm waves in the North Atlantic commonly have a length of 400 feet and a height of 20 feet, but in times of exceptional storms they may attain a length of over 1000 feet and a height of more than 40 feet. Captain Tanner of the United States Navy has photographed a storm wave exceeding 50 feet in height. The velocity of the waves increases with their length and is about 25 knots, for example, when the waves are 400 feet long and 40 knots for those 1000 feet long.

When great waves run out of the storm they decrease in height and become more rounded, but they may continue for hundreds or thousands of miles as the long heavy undulations of the surface known as " *ground swells.*" In this form the waves may attain a length exceeding 2000 feet.

The wave motion decreases rapidly in depth as shown in Fig. 126 and, according to theory, the orbit of movement of a particle at a depth of one wave length should be only 1/534 of that at the surface. That is to say, in storm waves 20 feet high and 600 feet long, the surface particles of water will move through orbits having a diameter of 20 feet, and those at a depth of 600 feet through an orbit only $\frac{2}{3}$ of an inch in diameter. If, on the other hand, the wave be 450 feet long and 30 feet high, the orbit of movement of a particle at 50 feet down would be 15 feet, at 100 feet down 7.5 feet, at 150 feet down 3.75 feet, and at 450 feet down about half an inch. Even great storm waves, therefore, cannot disturb the bottom below a depth of a few hundred feet. Strange as it may seem, the depth to which the wave is effective depends rather on its length than its height. Observations on the depth to which the sea floor is affected by waves are rather conflicting, undoubtedly because the depth affected varies from place to place, depending on several factors. Observations by divers off the south coast of England have shown that pebbles are rolled about at a depth of 50 feet during a storm, and Fol

records that, at a depth of 100 feet, he was tossed back and forth by the oscillation of the water on the bottom when the ground swells were running at the surface. Cobbles weighing a pound or more are sometimes washed into the lobster pots at a depth of 180 feet off Land's End, England, and wave ripples have been recorded in very fine sand at a depth of 617 feet near Madagascar. The depth at which fine, soft mud reposes on the sea floor may be considered as a good test of the depth of wave disturbance, since mud could hardly remain per-

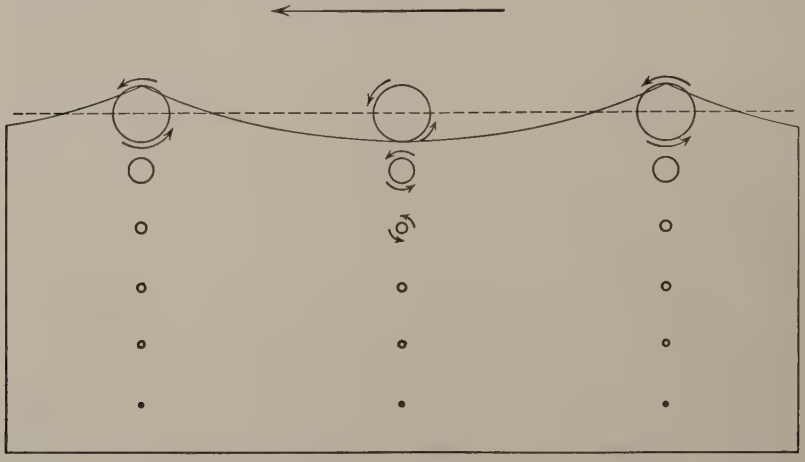

Fig. 126. — To show the decrease of wave motion with depth. The circles show the relative size of the orbits of motion at their respective depths as the wave passes. According to theory the diameter of the orbit of movement of a particle is reduced by one-half for each successive increase of one-ninth a wave length in depth. (After Johnson.)

manently on a wave-disturbed bottom. Judged by this criterion, the sea floors off exposed coasts are generally affected by waves to a depth of 200 or 300 feet, and exceptional waves move fine sediments to the edge of the continental shelf, a depth of about 600 feet.

When a wave passes into shallow water, an important change takes place in its form and behavior. The wave becomes higher and shorter and its front side steeper and more deeply concave until the crest arches forward, unsupported, and collapses with a roar. These collapsing waves or *breakers* form the *surf*, so common a feature along coasts. The height of the wave determines the depth at which it will break, and thus small waves break in shallow water, near shore, whereas the great storm waves break where the water is from 10 to 20 feet deep.

Tides. — Due to the attraction of the moon, modified by that of the sun, the ocean surface is lifted into two vast but low tidal bulges, one on each side of the Earth. These bulges remain fixed with respect to a

line extending to the moon, but since the Earth in its daily rotation turns under them from west to east they seem to run round the Earth from east to west.

In the open ocean, the surface merely rises a little and then subsides again as each tidal bulge passes; but where the bulge impinges against a coast the water is dragged forward, piling up on the shore and then receding, thus producing the familiar phenomenon of the *tides*. The height of the tide is determined largely by the configuration of the coast.

Fig. 127. — High and low tide in the Bay of Fundy. Note the same bridge and sailing vessel in both views. Port Williams, Nova Scotia.

On open, exposed coasts it is not more than 6 or 8 feet, and in protected embayments as, for example, the Gulf of Mexico, it is only 1 or 2 feet; but in estuaries that open out toward the advancing tide, the water rises higher as it converges, bringing about exceptional conditions like those in the Bay of Fundy, where the tide rises normally 30 to 40 feet and exceptionally as much as 50 feet. (Fig. 127.)

The immense bodies of water thus moving in and out of bays and estuaries and along the coast every 6 hours, produce strong tidal currents, which, like rivers, have a two-fold geologic function in that they both erode and transport. The work done by the tides in scouring the bottom and transporting material along some coasts is very great. The rise and fall of the tides indirectly aids in the attack of the waves on the land by increasing the vertical range of their contact with the lands.

Currents. — There are several types of marine currents, only the chief of which need be considered here. *Surface currents* are produced by the friction of the wind when it blows persistently in one direction for a time. Where they strike the coast and, with the aid of the breaking waves, tend to pile the water up on the shore, there is a compensating bottom current, the *undertow*, flowing seaward (Fig. 128). On a straight

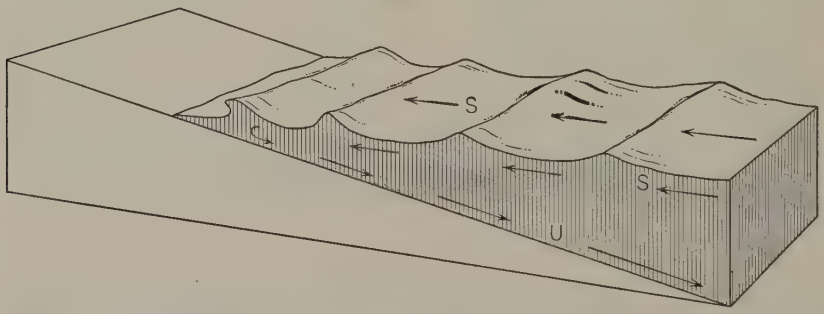

Fig. 128. — To show the relation of the wave-formed surface current (*S*) and the undertow (*U*) when the waves are coming directly onshore.

coast, with a gently shelving sea floor, the undertow is so diffused as to be very gentle, but where irregularities in the bottom constrict the returning current to more definite channels, the undertow may be a menace to swimmers and a powerful agent in sweeping sediments seaward into deeper water. Where the waves strike the shore obliquely, the run of the water due to successive impulses of the waves generates a current parallel to the shore, known as the *littoral* or *shore* current (Fig. 129). Such currents sweeping material along the coast are responsible for the formation of beaches, spits, and bars. The fine sand of Daytona Beach, Florida, for example, comes not from the interior of that limestone country, but along the shore from far to the north, where the rivers of Georgia and the Carolinas have brought it down to the sea out of the southern Appalachians.

Tidal currents are strong where the tide sweeps in and out of deep estuaries or bays. Where the shoreline is oblique to the east-west path

of the tidal bulge, there may be a strong component of the tidal current running parallel to the shore. This is true, for example, along the western coast of Newfoundland, where the ebbing tide sets to the northeast through the Strait of Belle Isle with a velocity of 4 or 5 miles per hour, a rate that compares well with the flow of large rivers.

The more general *ocean currents* have for the most part only an indirect geologic effect by modifying the climate of the lands, for they

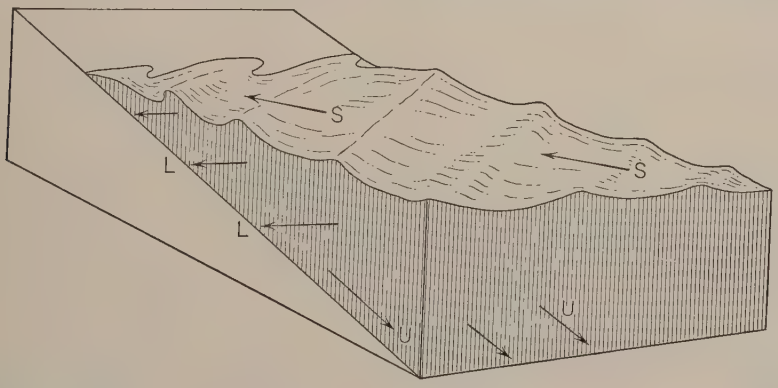

Fig. 129. — To show the relation of the wave-formed surface current (*S*) to the littoral current (*L*) and the undertow (*U*) when the waves are oblique to the shore.

rarely touch the shores or the bottom with sufficient velocity to erode or transport sediments. On the other hand, they have a distinct geographic value in drifting organic waifs to strange lands. These currents are generated by several different factors. The unequal heating of the ocean by the sun in tropical and polar regions would establish a slow general circulation of its waters through convective movements. This action, however, is controlled, hastened, and magnified by the wind belts of the Earth, described in Chapter III, and by the disposition of land and sea. Driven by the trade winds, there is, in the equatorial regions of the Atlantic as well as the Pacific, a broad current moving westward along the surface. When this strikes the continental coasts it divides, one part turning northward, the other southward; and each, circling, returns to the equatorial belt, thus making in each ocean a vast eddy, one north, the other south of the equator (Fig. 130). In the same manner there is a circling movement in the Indian Ocean. In the center of each ocean are more stagnant areas, that in the North Atlantic being known as the Sargasso Sea. When these broad slow movements, which are known as *drifts*, approach the coasts, the water tends to accumulate, and where confined by the configuration of the land is hastened in its motion, giving rise to *streams*. Thus in the North

Atlantic, the equatorial current striking the north coast of South America is deflected northward. A part enters the Caribbean Sea and the Gulf of Mexico, whence it issues through the greatly confined Straits of Florida and passes northeast into the Atlantic as the well-known Gulf Stream. In the straits this current averages 72 miles a day and during the summer and winter months sometimes rises to 120 miles a day; but as it spreads and approaches mid-ocean the velocity diminishes greatly, falling finally to 10 miles a day.

As this drift approaches the shores of Europe, it divides and one part turns southward to pass along the coast of Africa, and so to rejoin the

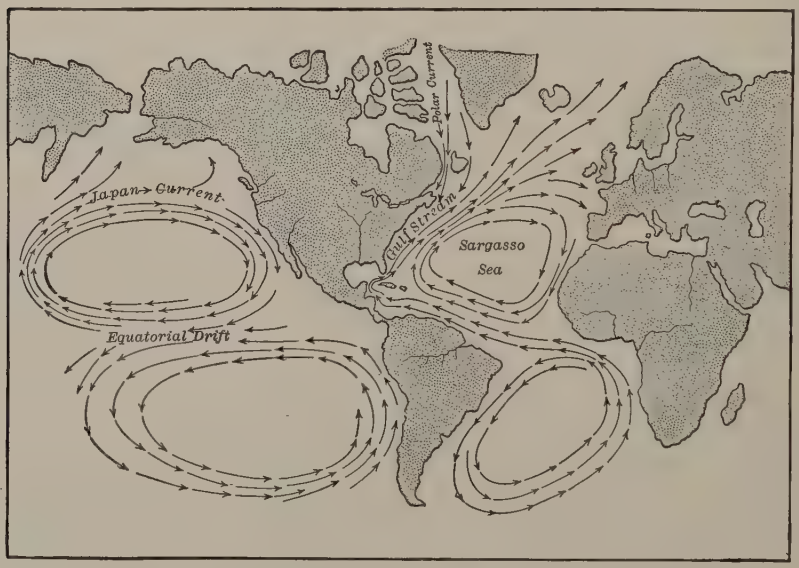

Fig. 130. — Map showing main ocean current and drifts.

westward equatorial drift. Another portion passes northward into the Arctic Ocean, and to balance this a cold current comes down from the western coast of Greenland, past Nova Scotia and New England, and gradually passes under the warm surface of the Gulf Stream to sink into the abyss.

In a similar way, in the North Pacific a current moves along the coast of Asia, then eastward and finally southward along the western coast of North America. It is known as the Japan current. These warm currents, moving into northern latitudes, have a great effect upon climatic conditions in the lands whose shores they strike. Thus the ocean currents, like the atmosphere, by taking part in the general circulation

on the surface of the globe are great distributors of heat. They carry warmth into the polar seas and, returning as cold currents, bring with them reduced temperatures and icebergs to be melted in warmer regions. Were it not for their agency, ice would continually increase in the polar areas. The indirect effect of the ocean currents on geological processes is therefore important, although these currents perform little direct geological work.

Differences in temperature may generate ocean currents, the denser, colder water tending to move toward the lowest places and to displace the warmer water toward the surface. For this reason there is a slow and general movement of the polar waters along the bottom into all the deeper ocean basins, which consequently have a temperature approximately at the freezing point of surface fresh water. In the depths of the Pacific, Indian, and South Atlantic, for example, the temperature ranges between 1° C. and 4° C. If there were no cold polar regions, the ocean bottoms would not be so cold.

Differences in salinity may also set up vertical currents, because the density of the water increases with its salinity. The lighter fresh water of many great rivers extends far out over the surface of the salt water, mingling gradually by diffusion. On the other hand, in regions of excessive surface evaporation the upper water of the ocean may become concentrated enough to form downward currents compensated by the rise of cold water from the depths. Professor T. C. Chamberlin developed the ingenious hypothesis that, at times in the geologic past, this factor may have reversed the present vertical circulation in the oceans. At present the influence of the temperature dominates and the general vertical circulation carries the cold waters down and equatorward, the warm waters moving poleward at the surface until their warmth is gradually lost by radiation. Chamberlin suggests, however, that at times of warmer polar regions the effect of salinity may have dominated, the evaporation in equatorial regions then causing the warm surface water to settle and be crowded poleward along the bottom, whence it emerged with its stored heat in high latitudes. In such an event, the polar regions would be much more effectively warmed and the tropical heat ameliorated by the surface flow of cooler water, the polar regions remaining shrouded in fog due to the ascending warm water. While this speculation offers a possible explanation of the more equable world climates of many past geologic ages, it must be confessed that there is no direct evidence that can be cited in its support.

Ascending currents bringing cold water from the depths near some of the tropical shores render the climate of the coastal belt surprisingly cool. This is true of the coast of Peru, and likewise of parts of the north-

west coast of Africa as, for example, the coast of Morocco which lies almost at the edge of the tropical desert but is refreshingly cool. It is partly because of the ascending currents along the Pacific coast of California that San Francisco and Los Angeles enjoy summer temperatures so far below that of Washington, D. C., and the Virginia Capes.

Marine Erosion

Erosive Processes. — The surface of the seas, like an ever-moving horizontal saw, is ceaselessly cutting and gnawing away at the lands. This it does in several ways, producing a variety of features that are worthy of consideration.

Sea water has more or less solvent effect on rocks, tending to disintegrate them, and thus aid in their destruction; but the chief factor in marine erosion is the mechanical attack of the waves. The force of impact of large waves due to the sheer weight of tons upon tons of surging water may be sufficient to erode unconsolidated sediments rapidly or to move enormous blocks of stone. According to theory, the pressure exerted by a wave 10 feet high and 100 feet long is 1675 pounds per square foot, and of a wave 12 feet high and 200 feet long 2436 pounds per square foot; and "great ocean waves, if we assume a height of 42 feet and a length of 500 feet, should produce a pressure of 6340 pounds per square foot" (Johnson). These theoretical calculations accord well with actual dynamometer measurements, which on the coast of Scotland showed the average force of the summer waves to be 611 pounds per square foot and the average for the winter months to be 2086 pounds with extremes up to 6083 pounds. The damage done by storms to harbors and breakwaters bears further testimony to the force of the waves. During a great storm at Wick, Scotland, in 1872, a solid mass of stone and concrete weighing 1350 tons was torn from its place at the end of a breakwater and dropped unbroken inside the pier. When great waves strike against a sea wall or solid cliff, they not uncommonly dash up to heights of 100 feet or more. At the lighthouse on Tillamook Rock, on the exposed coast of Oregon, the water of the waves was thrown over 200 feet above the sea during a storm in 1902, and during the winters of both 1912 and 1913 the impact of the waves broke panes of plate glass in the lantern of the same lighthouse at a height of 132 feet above the sea. It is not surprising, therefore, that the faces of bold cliffs are shattered, and softer materials rapidly dislodged, simply by the impact of the waves. (Fig. 131.)

Hydraulic pressure may be a powerful agent of the waves as well, for most rock masses have crevices or larger cracks in them, and the

air or water in these, driven violently in by the impact of the waves, acts as a wedge, disrupting them and dislodging large pieces. In this way heavy masonry is often torn asunder. Moreover the water, rushing into cavities and suddenly retreating, leaves a partial vacuum which tends to suck away portions of the roof and sides, and the constant repetition of this action aids in forming sea caves, blowing holes, and spouting rocks, so frequently seen on rocky coasts.

The chief eroding action, however, is accomplished by abrasion, and in performing this work the waves use as tools the dislodged material,

Fig. 131. — Storm waves breaking against the sea wall at Hastings, England. (After Johnson. Photo by Judges.)

and also that which falls from above and tends to form a talus at the foot of the sea cliff. The constant striking and grinding, not only of sand and gravel, but even of heavy boulders, render the waves formidable agents of destruction, through whose work even the hardest rocks are worn away. The ineffectiveness of clear water, aside from the mechanically disrupting processes, is strikingly shown in places on the coast of Norway, where headlands extend into water too deep to be affected by coastal débris. The rock surfaces, smoothed and furrowed by former action of glacial ice, still retain these characteristic features, though subjected to the constant washing of the waves.

In the process of abrasion the material used by the waves is itself ground up and reduced to sand and silt. It loses its angular character and takes on the rounded form characteristic of coastal débris submitted to chafing by the waves.

Transportation of Sediment. — If wave erosion is to continue, the ground-up rock débris must be removed, like sawdust from the track of the saw, in order that fresh rock surfaces may be exposed to attack; otherwise the fine material would act as a buffer to receive the waves and prevent further erosion. This removal is done by the undertow, which takes the débris out to sea, and also by tidal and littoral currents, which carry it away. Were it not for the aid of these agencies, wave erosion would cease except where the sea encroaches on a sinking land mass.

As in the case of streams, the products of erosion may be transported in the sea, either in *solution*, in *suspension*, or by *bottom rolling*. But unlike stream transportation, the currents in the sea are aided greatly by the waves.

Due to the orbital motion of the water in an ordinary wave of oscillation, a grain of sediment on the bottom tends to be lifted and carried slightly forward as each wave crest passes, but is carried down and back to its original position by the trough of the wave, so that no actual transportation is accomplished by the wave itself. But when the particle of sediment is lifted free of the bottom, even the gentlest current can deflect its fall; in this way gentle currents, aided by the repeated lift of the waves, can sweep along sediments that they alone would be powerless to move.

When waves reach shallow water and begin to drag heavily on the bottom, their movement ceases to be one strictly of oscillation, for the water then tends to roll ahead as a wave of translation which, like a strong current, drags the bottom material forward with it. Since the waves drag bottom only when approaching the shore, they tend to transport the bottom sediment landward. This tendency is opposed, however, by the undertow and by gravity, both tending to shift the sediment down the seaward slope. These seaward forces act continuously and are able, therefore, to overcome the stronger but intermittent surges of the waves. Although the sediment is dragged back and forth with each passing wave, the net movement of the fine material is seaward. On the other hand, the undertow may be unable to transport coarse material which the breaking waves drag forward, and when this is true the coarse sediment migrates landward even though the fine material is being shifted seaward. Murray has recorded, for example, that stone ballast discharged from ships near the British coast in water as much as 60 feet deep, has been thrown up on the beach during a subsequent storm. This factor keeps the coarse sediment concentrated at the shore, where it is known as the *beach shingle*.

Even fine sand is thrown shoreward to form a sandy beach where

loose sediment is sufficiently abundant, the waves carrying forward more of the material than the undertow can remove. The beach is but a temporary accumulation of sediment in transit, however, for there is a constant loss to the undertow and shore currents that spread the material seaward.

Physiographic Development of the Shore

In their attack on the shore the waves and currents carve erosive features in some places and in others build up deposits of the loose sediment. The nature of the shore features thus produced at a given place depends partly upon the character of the land surface against which the marine agents have to operate and partly upon the length of time they have been at work. As a land surface under the influence of stream erosion passes through a cycle of physiographic changes from youth to maturity and old age, so also the shoreline passes through a predictable series of changes as the marine cycle progresses, and many of the shore features, like those of a degrading land surface, are temporary features in the physiographic cycle.

The marine cycle is inaugurated ordinarily in one of two ways; either (1) by an uplift of the sea floor against which the waves start their attack *de novo* or (2) by a submergence of the coast which brings the sea into contact with a former land surface. The initial form of the shore on an emergent sea floor is utterly different from that of submerged land and the sequential forms developed from the one are so distinct from those of the other that each type of shore must be described separately.

The Shoreline of Emergence. — When a sea floor emerges by gentle uplift it forms a nearly flat *coastal plain* mantled by unconsolidated sediments. The plan of this shoreline is at first very simple, since it marks the intersection of this flat surface by sea level. The water is so shallow for a considerable distance from shore that the waves drag heavily upon the bottom, picking up the loose sand and gravel, which they carry forward to the line of breakers. Here the loose material is dropped again as the wave spends itself, gradually building up a narrow submarine ridge parallel to the shore and just within the line of breakers. Storm waves eventually build the deposit above sea level, forming a long low sandy island parallel to shore (Fig. 132, A). This structure is known as an *offshore bar* or *barrier beach*. The narrow body of shallow water lying between the offshore bar and the shore is a *lagoon*. Large lagoons are also called *sounds* but not all sounds are lagoons. Ordinarily the sweep of the tide in and out of the lagoon keeps inlets

through the barrier open to the sea. The Atlantic and Gulf coasts of the United States display these features in grand development (Fig. 143). The width of the lagoon depends on the seaward slope of the bottom and the size of the storm waves. It is commonly a mile or more and may be several miles. At Cape Hatteras the offshore bar is about 20 miles from the mainland but along the east coast of Florida it is near shore and the lagoon is known as Indian River. The city of Galveston is built on an offshore bar and its tragic flood of 1900 occurred when high seas, driven by a hurricane, rose 15 feet above their normal level.

The sea floor outside the barrier is gradually excavated by the waves, supplying the material of which the barrier is made. As its depth in-

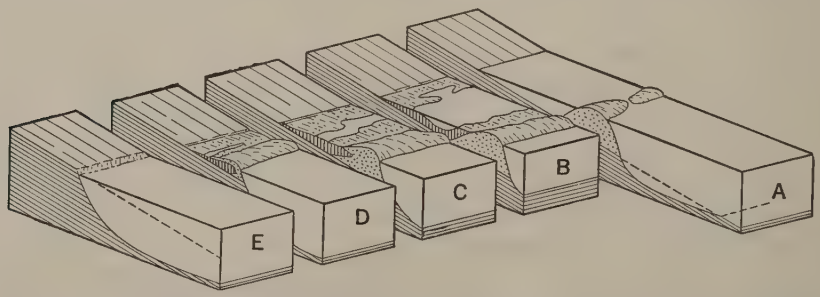

Fig. 132. — Showing youthful stages in the development of a shoreline of emergence. The offshore bar is stippled in section and the lagoon filling is vertically lined. *A*, early youth, showing the formation of the offshore bar; *B*, a later stage showing partial filling of the lagoon; *C*, a still later stage in which the offshore bar has been prograded toward the shore and the lagoon has become a fresh-water marsh; *D*, late youth; *E*, early maturity, the offshore bar and lagoonal deposits having disappeared. Vertical scale exaggerated. (After Davis.)

creases the waves are less impeded by friction on the bottom and are finally able to break with force against the barrier itself, eroding material from its seaward side. A part of this débris is carried out to sea, and during storms a part of it is thrown over the bar into the edge of the lagoon. In this way the barrier is gradually prograded shoreward. Meanwhile the lagoon, rarely more than 20 feet deep at the start, tends to be filled by sediment washed into it from the land, as well as by that thrown over the barrier. Where the water is shallow enough, vegetation thrives and adds its quota to the accumulating sediment (Fig. 132, B). As the barrier is gradually driven shoreward the lagoon becomes a salt-water marsh with open tidal channels (Fig. 132, C and D). Eventually both barrier and lagoonal deposits are completely cut away as the sea bottom is excavated up to the shoreline (Fig. 132, E) and then for the first time the waves begin their attack upon the land.

During these early stages of the cycle, the bottom profile near shore

has been gradually changing from the simple slope of the original sea floor to the compound curve represented in Fig. 133. There is a concave portion near shore where the bottom is being eroded. It extends seaward as a nearly flat *wave-cut bench* or *terrace* to the line of depth where wave cutting is ineffective. The depth of the wave-cut terrace depends on the size of the prevailing storm waves and the supply of sediment they have to move, but it is generally between 10 and 20 feet near shore and gradually deepens seaward through abrasion by the sediment that is shifted across it. Beyond the zone of cutting the sediment is dropped and gradually built out as a submarine embankment or *wave-built terrace*, which continues the nearly flat slope of the wave-cut terrace.

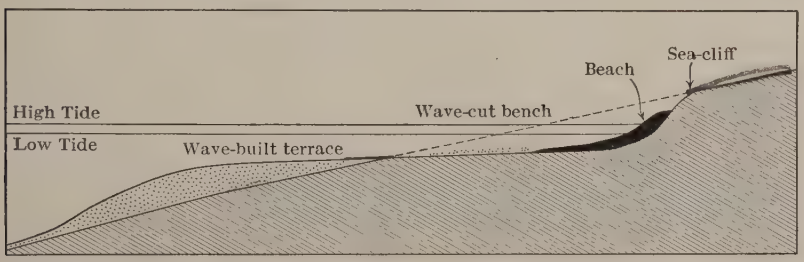

Fig. 133. — The wave-formed shore profile of equilibrium, concave near the shore where the sea is cutting and convex farther out where sediments are accumulating. The beach is represented by solid black and the sediments of the wave-built terrace are stippled. Vertical scale exaggerated.

The bottom is convex over the outer part of this deposit and concave further out where the sea floor drops away to the normal bottom slope. The curve shown in Fig. 133 is the *profile of equilibrium* and its general character will be maintained to the end of the marine cycle, though the curve will gradually flatten out as the sea cuts farther inland and the waves waste more and more of their energy in bottom friction while crossing the wide neritic zone.

Where the waves are cutting into solid rock the wave-cut terrace is rock-floored, though it may be more or less mantled with débris in transit from the shore; but where the attack is upon unconsolidated sediments the wave-cut terrace obviously has a floor of loose material.

As the waves concentrate their attack at sea level, cutting horizontally into the sloping land surface, they tend to undercut the land. The overhanging material is gradually dislodged and falls down and thus the coast comes to be terminated in a *sea cliff* as in Fig. 132, E. The nature of the sea cliff depends in part on the material attacked by the waves. In firm rocks it may be very steep or even overhanging, but in loose material like sand, slumping and sliding keeps pace with

undermining and the sea cliff is a slope equal to the angle of repose of the loose material. The form of the sea cliff is also influenced by the relative rate at which the shore material yields to weathering and to wave cutting. For example, clay is easily cut by the waves and commonly presents bold bluffs in spite of its softness, whereas hard granite may succumb more rapidly to weathering than to the attack of the waves, especially where it is much jointed and exposed to frostwork. The hard rock may thus present a sea front of low slope whereas soft rock may form cliffs.

The débris that falls from the sea cliff or that is eroded at sea level (as well as that which is introduced by streams) is shifted back and forth by the waves and currents and the finer material is carried seaward with the undertow. The coarser sediment is kept at the shoreline, however, to form the *beach*, a low ridge of sediment thrown up by the waves along the shore. As indicated in Fig. 134, the beach is only a thin veneer on the bedrock floor, extending landward to the limit reached by the greatest storm waves. In cross section it is convex upward at the shore, where it is thickest. It extends for a short distance under the water, thinning gradually and merging into the sheet of sediment that partly mantles the rock-cut bench while in transit out to sea. The upper margin of the beach is generally marked by a belt of coarser material thrown up by the heaviest storm waves and this grades into finer material toward its seaward margin. In places where the shore zone is formed wholly of fine sediment the beach is made of sand alone.

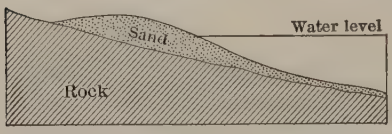

Fig. 134. — Section of a beach. (After Gilbert.)

If the rock forming the sea cliff is cut by vertical joints, the waves tend to widen the joints and quarry out some of the blocks, leaving others standing isolated as *stacks* (Fig. 135). If the joints are more irregular and intersect before reaching the summit of the cliff, *sea caves* are formed at the base of the cliff. As the waves break into the mouth of a sea cave they exert a heavy hydraulic pressure on its sides and roof, followed by rarefaction of the air as the wave recedes. By this action the roof of the cave may be excavated through to the surface so that spray will dash up through the roof with each breaking wave. Such caves are known as *blow holes* or *spouting caves*. Narrow promontories or islands are commonly undercut by the development of sea caves to form *sea arches*.

Obviously the greatest inroads are made by the sea where the rocks are weak and the slightest progress where they are resistant. For this

Fig. 135. — Sea cliff and stack, the latter a remnant of former land now eroded away. Coast of Wales. (Geol. Surv. of England and Wales.)

Fig. 136. — A bay with a curving beach. Conception Bay, Newfoundland. (Walcott.)

reason a simple shoreline formed of heterogeneous rocks gradually develops irregularities as erosion progresses; the resistant masses stand out to form *headlands* or *promontories* and the weak rocks recede to form *coves* or *bays* (Fig. 136). If the rocks are composed of parallel layers or beds and their edges are exposed to the waves, the weaker, softer layers are rapidly worn away and the harder beds, left unsupported, break away in blocks. If the beds are horizontal the harder layers may project for a time as table rocks with cavities under them, as on the coast of Lake Superior. If the beds are vertical, or inclined, but with edges exposed, the hard layers stand out like columns or ribs (Fig. 137). If

Fig. 137. — Irregular coast and sea cliff produced by erosion in nearly vertical rock strata. Pembroke, Wales. (Geol. Surv. of England and Wales.)

the face of the beds is towards the sea, erosion is slower because the hard layers form an apron, or wall, to protect the soft layers behind them. If the rock masses are homogeneous and hard, like trap or granite, the irregularities are largely determined by joints and by the arrangement of these with respect to the sea front. Thus a bold coast facing the sea is likely to show many minor irregularities of topography.

As one notes the delicate adjustment of the shore configuration to the resistance of the rocks he might be inclined to think that all the irregularities of the shoreline are due to erosion. Such, however, is far from true for there is an effective limit to the depth inland to which embay-

ments may be cut by the waves. For example, the deeper the bay the more completely the defending headlands protect it from all oblique waves. Experience shows that large storm waves do not enter far into deep embayments. Moreover the débris torn from the headlands tends to be swept into the protected bay-heads to form beaches that mantle and protect the shore. The great and deep embayments of the coast, therefore, result not from marine erosion but from crustal warping or the drowning of river valleys.

In the physiographic development of the shoreline of emergence, the stages illustrated in Fig. 132 are all preliminary to the real attack against the land. They represent the *youthful stage* of the marine cycle. Following the disappearance of the offshore bar, the sea launches its drive with full vigor against the shore. This is the beginning of the *stage of maturity.* The sea cliff is now formed and the simple shoreline rapidly develops irregularities, until the plan of the shore is adjusted to the forces of attack and to the resistance of the rocks. Erosion then proceeds gradually along the whole line. The final stage of *old age* is attained only when the sea has cut so far inland that the waves spend most of their force in friction over the shallow bottom and make but a feeble attack upon the shore. After this stage is reached, further retreat of the shore due to marine erosion proceeds very slowly.

The Shoreline of Submergence. — The initial form of a shoreline of submergence is marked by extreme irregularity, since the marine water comes to rest against an eroded land surface, entering far up the valleys and forming deep irregular embayments between sloping headlands. Glaciated valleys thus drowned become *fiords* and the lower courses of the normal stream valleys become *estuaries* or deep bays. Hard-rock islands are likely to be abundant near the shoreline, representing isolated hills of the drowned land surface.

The *youthful stage* of the cycle of erosion of such a shoreline is marked by lack of adjustment of the erosive forces to the plan of the shoreline. The waves must concentrate their attack chiefly on the offshore islands and the headlands that defend the deeper embayments. These are quickly cliffed, while the bay-heads tend to be silted up by the streams that enter them. In their fresh attack against the land surface, the waves quickly indent the weaker spots, developing *sea caves, arches,* and *stacks* and forming little bays even in the headlands. The irregularities thus developed are on a small scale and are only incidents in the general process of straightening the shoreline, but they give to it a crenulate plan characteristic of early youth.

Three youthful stages and one of early maturity of the submergent shoreline are indicated in Fig. 138. It will be noted that they tend toward

Fig. 138. — To show early stages in the development of a shoreline of submergence. *A*, early youth, showing cliffed islands and headlands without beaches; *B*, a later stage showing recession of the headlands and the development of winged headlands through the growth of spits and hooks; *C*, later youth showing truncated headlands united by bars and extensive salt-water marshes in the heads of the bays; *D*, early maturity. The scale of this figure is much smaller than that of Fig. 132, so that each block represents a larger area than in that diagram. (Modified from Johnson.)

a simplification of the extreme irregularities of the initial shore. The first stage is marked by *cliffed headlands*. Since the surface is drowned, deep water comes close to the headlands and the débris first eroded sinks beyond reach of the waves so that no beach is produced. Gradually the headlands are cut back and the débris supplied by their erosion accumulates until a deposit is built up into the zone of wave action. A beach is then formed at the foot of the cliff. Littoral currents do not enter deeply into the estuaries and as they sweep past the truncated ends of the exposed headlands they carry part of the beach material laterally into deeper water where it settles, building out a submarine embankment. By continued action of this process the embankment is

Fig. 139. — A curved spit, or hook. Duck Point, Grand Traverse Bay, Lake Michigan. The left end, beyond the edge of the view, is attached to a point of land. (U. S. Geol. Surv.)

built out across the adjacent embayment very much as an artificial fill is made across a valley by the dumping of dirt at the end of the fill. Incoming waves will throw some of this material upon the embankment and build it above sea level. It then appears as a low sandy or gravelly promontory built out from the beach. If nearly straight this structure is a *spit*. Another current crossing the end of a spit may sweep the material aside, developing a curved or hooked spit which is known as a *hook* (Fig. 139). If the deposit is extended until it closes, or nearly closes, the mouth of an embayment it is known as a *bar* (Fig. 140). Spits and hooks tend to form wherever the littoral currents sweep sediment away from the beach, whether it be at the end of a promontory or merely of a blunt cape. Bars tend also to form in the sheltered water between

islands and the shore, resulting in *land-tied islands* known as *tombolos* (Fig. 141). *Sea caves*, *arches*, and *stacks* are especially characteristic of these early youthful stages of the submergent shoreline.

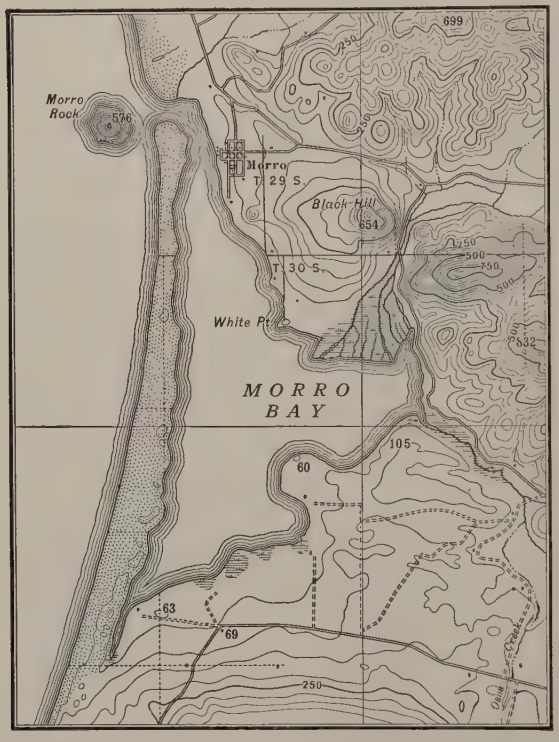

Fig. 140. — A bar closing Morro Bay, Calif. (U. S. Geol. Surv.)

While spits, hooks, and bars are forming about the retreating headlands the inner ends of the embayments tend to be filled by deltas where streams enter these protected bodies of deep water.

Eventually, as in Fig. 138, C, the shoreline is much simplified as the land-tied islands are cut away and the headlands truncated and united by more or less continuous bars. The bay-heads become largely filled with sediment and pass into salt-water marshes like those that are so common along the New England coast at the present time.

When finally the headlands have been completely cut away and the bay-head fillings are all removed, as in Fig. 138, D, the *youthful stage* of submergent coast is completed and *maturity* has begun. The waves are now at work along the entire coast, the plan of the shoreline is adjusted to the resistance of its rock masses, and the bottom has developed the *profile of equilibrium*, concave over the wave-cut bench and convex over the wave-built terrace. The deeper irregularities in the sea floor have been filled with sediment. From this stage on the development is identical with that of the shoreline of emergence.

Shorelines Produced by Interrupted Marine Cycles. — Completion of the marine cycle would require stability of the land and a constant sea level for a very long time. A relative uplift or depression of the coast will interrupt the cycle and superpose a new cycle on the old.

Fig. 141. — A tombolo or land-tied island. Bay of Fundy, Nova Scotia. The left side of the connecting bar is not fully shown.

Fig. 142. — An elevated wave-cut terrace north of Port Harford, Calif. Old stacks rise above the terrace. (U. S. Geol. Surv.)

If uplift of the coast is intermittent, the wave-cut bench formed during a period of stationary sea level may be lifted above the sea to form a *marine terrace* paralleling the shore (Fig. 142). Such terraces are strikingly developed for long distances along the southwest coast of Newfoundland. At Port au Port, one of them bevels evenly across the edges of steeply tilted strata and attains a width of more than 2 miles. Although originally carved by the waves, it is now 55 feet above sea level. Abundant marine shells occur in the clay that mantles another terrace lying just 100 feet above sea level along the western end of Cape St. George, Newfoundland. Farther north in both Newfoundland and Labrador, there are remnants of older and more eroded marine terraces at several elevations up to about 500 feet above the present sea. Along the coast of California, marine terraces have been recognized up to 1600 feet above sea level.

Fig. 143. — Map of the coast of North Carolina showing an association of drowned valleys, wide lagoons, and offshore bars.

The landward margins of some marine terraces are marked by ancient sea cliffs accompanied by other shore features such as stacks and sea caves. Such uplifted sea caves are used as shelters by inhabitants of the bold coast of Fifeshire in Scotland (Fig. 197). Characteristic beach deposits also mark the position of some old shorelines, and where uplift has been intermittent, a series of beaches may lie one above another as they do along the coast of southeastern Labrador, where the land has only recently recovered from its depression by the weight of the Pleistocene ice cap.

The dead, blanched forms of uplifted coral reefs also bear mute testimony to a relative uplift of certain tropical shores.

Coasts frequently undergo uplift followed by submergence, or depression followed by later uplift. The result of such oscillations in the level of land and sea is the production of a variety of shore features some of which have resulted from emergence and others from submergence. Thus, for example, the Central Atlantic coast of the United States (Fig. 143) is marked by the drowned valleys characteristic of a

submergent shoreline and by the offshore bars characteristic of an emergent coast. An extensive submergence has been followed by a recent but lesser emergence, leaving the valleys drowned and at the same time reducing the depth of the near-shore water to such an extent that barrier beaches have been developed. Under these conditions the barrier beach at first has extensive interruptions marked by the submerged valleys. The barrier beach may be extended across these depressions as a result of lateral shift of sediments by littoral currents, growth of the barrier beach taking place by the development of subsidiary spits and hooks and bars. There are good examples of this relationship along the south shore of Long Island.

Some coasts show old, dissected marine terraces far above sea level, bearing witness to an earlier uplift of the land, whereas the present shoreline is indented by drowned valleys as a result of more recent submergence.

Distribution of Marine Sediments

Shallow-Water Terrigenous Deposits. — The land-derived (terrigenous) marine sediments all have as their goal the quiet depths of the sea or ocean bottom, their temporary deposition in the form of beaches, bars, and barriers being merely an incident in their seaward journey. Constant agitation by the waves during their transportation imparts to the marine deposits a much higher degree of sorting than is usually attained by river-laid sediments, a fact suggested by the cleanness of fine beach sand.

Where wave and current action is vigorous, only coarse material can accumulate, but as this activity declines, finer and finer sediment comes to rest. The coarse sediments are therefore deposited in shallow water, and, in general, the deposits grow finer of grain with increasing depth. But since very shallow sea bottoms may be protected by sheltering land masses or by coral reefs, it must not be supposed that fine-grained sediments are all of deep water.

In the neritic zone of the sea the sediments vary greatly in character from place to place. For example, along the bold and exposed coast of California the near-shore sediments include gravel and coarse sand; off the mouth of the Mississippi the deposits are finer sand and mud derived from distant highlands; and the low limestone coast of southern Florida is bordered directly by deposits of the finest limy ooze.

Soundings show that modern gravels are generally confined to depths of less than 30 feet and are deposited within 1 mile of shore, though exceptionally gravel is carried to depths of 150 feet and to distances up to 10 miles from shore. Sand generally does not extend beyond a depth

of 250 or 300 feet and the outer portion of the continental shelf is therefore covered by muds alone.

Coral Reefs and Calcareous Deposits. — In the shallow water along many tropical and subtropical coasts coral reefs play an important geological rôle. These limy reefs are estimated to have a combined area of half a million square miles at the present time and the calcareous débris produced through their erosion by the waves is spread over a much greater area of the sea and ocean floors. The earlier observers were so impressed by the limy deposits of this nature that they attributed limestones in general to a coralline origin. It is now realized, however, that other shell-forming animals, as well as lime-secreting plants (calcareous algae), play a large part in the formation of coral reefs and, moreover, that extensive limy deposits are forming where reefs are unimportant. Nevertheless the reef-making corals still hold an exceptional geologic interest.

Corals are small marine animals of low organization characterized, among other things, by the fact that they secrete an external limy skeleton about the base of the body. The name *stony coral* is commonly applied to the skeleton in distinction from the living animals which are called *polyps*. Corals reproduce by forming daughter polyps which bud from the margin of the parent in plant-like fashion and retain a fleshy connection, so that living colonies of many thousands of individuals combine in the secretion of an elaborate limy structure of considerable size. Some of the colonies are massive and hemispherical or irregular in their growth, whereas others are branching and shrub-like. Some individual colonies reach a diameter of 15 feet or over and house a million or more little polyps, each living in its own tiny protecting chamber like a cliff-dweller in a city of stone (Fig. 144).

Where the reef corals thrive, there congregate many other forms of life, some of which also secrete limy deposits. Certain plants (the nullipores and other kinds of lime-secreting algae) add their share of extensive, encrusting layers of limy material; and in some reefs they predominate over the corals.

As the corals expand and build upward the older polyps die, and gradually the dead base is buried by the growth of other animals and by the abundant débris which storm waves break from the living parts above. This accumulating deposit is eventually cemented into a white cavernous limestone, upon whose upper surface the living corals grow and flourish. In this way the *coral reef* is formed. It rises until its surface is just below sea level, where it is laid bare only at the lowest tides for short periods, for corals can stand exposure to the air but a few hours at a time. Over the reef the waves swirl and break, and at its outer edges

the reef-building corals thrive best, for in the rush and dash of the waves they find the most food and calcium carbonate in the water, and the richest supply of life-giving oxygen.

Coral reefs are sharply limited in their distribution by the depth and the temperature of the water. Although individual corals of various kinds live in more varied environments, the reef-forming kinds thrive

Fig. 144. — Growing corals, seen at low tide. The view is from a barrier reef across the lagoon toward mainland 20 miles away. Great Barrier Reef, Australia. (Saville Kent.)

only where the water is clear, normally saline, shallow, and warm. They cannot endure a temperature below 68° F. and they live normally where the water is less than 150 feet deep, though rarely, where there are descending warm currents, they exist well below this limit. Coral reefs are therefore confined to the shallow waters of the tropics and near tropics, except where (as at Bermuda) warm currents form exceptional conditions. The distribution of fossil coral reefs in the older geologic deposits is consequently thought to carry special climatic significance.

According to their position and form, coral reefs have been grouped into three general classes: fringing reefs, barrier reefs, and atolls.

Fringing reefs lie close against the shore and form a bench or platform extending out toward the sea, laid bare only at very low tide. As the

corals cannot grow above sea level the growth of this platform is chiefly seaward.

The width of the reef seems to depend largely on the steepness of the land slope. Since the corals grow and flourish chiefly on the outer edge of the reef, the seaward slope of the latter is usually very steep. As material is broken off by the waves and rolls down the slope, it forms a rising talus which gradually becomes compacted and cemented by calcium carbonate, eventually forming a base upon which the corals advance the reef seaward. If the bottom slope is steep, the débris settles into deep water and much upbuilding is required to give a support for a small advance; if the slope is gentle the reef builds out more rapidly and may attain a width of a few miles. Opposite the mouths of streams the reefs are nearly always wanting because the corals cannot endure the freshened and muddy water.

Barrier reefs differ from those just described in that they are situated some distance from the land, and are separated from the shore by a *lagoon* of shallow water. Many of the high volcanic islands of the Pacific as well as the islands of the Caribbean region are more or less completely girdled by such encircling reefs. The west coast of the island of New Caledonia has a reef of this sort that extends for 400 miles, and the greatest of all coral reefs is the Great Barrier Reef which stretches for 1200 miles along the eastern coast of Australia, with an average distance of 20 to 30 miles from shore and a depth of 100 to 300 feet of water in its great lagoon.

A barrier reef may be from 1 to 30 miles from the shore and the average of the maximum depths of the lagoons in the Pacific is about 200 feet, though many of them are much shallower. Openings or breaks occur in the barrier because of the ebb and flow of the tide into the lagoons, and where they are sufficiently deep the lagoons serve as harbors.

Atolls are more or less ring-shaped reefs enclosing circular lagoons instead of islands (Fig. 145). The breadth of such rings may be but a fraction of a mile or it may be as much as 20 or even 50 miles, and the depth of water in the lagoons ranges from a few feet up to 300 feet. On the outside the slope may descend quite sharply thousands of feet to the ocean floor. Generally there are openings in the reef on the leeward side, which afford access to the lagoon.

Atolls, like either of the other types of reefs, may support low islands where storm waves have tossed the coral débris upon the reef and built it above sea level. The islands thus made are usually not more than 10 or 15 feet above normal sea level and from a quarter to half a mile wide, though often they are long in the direction of the reef. Most of them are covered by vegetation, including palms, and form spots of

great beauty. Some are inhabited, though their low elevation subjects them to the danger of being swept by the sea during exceptional storms. The only living atoll in America is small Sombrero of the West Indies.

Fig. 145. — An atoll. (After Dana, from an old picture.)

The Origin of Barrier and Atoll Reefs. — The origin of fringing reefs is easily understood. Since corals prefer shallow, warm water they tend to attach themselves near the shore and grow at first in scattered colonies which later coalesce as they increase in size. Thus a continuous coral growth is built up to low tide level, after which it can only grow outward as a fringing platform. It is equally clear that typical fringing, barrier, and atoll reefs are stages in a completely intergrading series. On the other hand the cause of the change from the fringing to the barrier or atoll form has given rise to much speculation, and the explanations proposed have invoked geologic changes of great significance. Three distinct theories may claim our attention.

The *subsidence theory*, first advanced by Charles Darwin, elaborated by Dana, and more recently rejuvenated by Davis, will be easily understood by reference to Fig. 146 (*a*), which represents a volcanic island that has slowly subsided. At an early stage, a fringing reef (*A*) has been formed. As explained above, the corals thrive best on the outward face of the reef. Accordingly, it may happen that after the reef has attained a considerable width only the more thrifty outer margin can grow fast enough to keep pace with the subsidence. The inner part then gradually becomes drowned and is covered by the lagoon behind the reef. As growth continues chiefly on the seaward face, the reef

grows outward farther and farther. Portions of the coral skeletons broken by the waves fall down to make a talus outside the reef and some of the finer detritus is washed over the reef into the lagoon, which thus tends to be silted up gradually with limy mud as its bottom subsides. Stage *B* in our figure represents this phase in the island's history, with the barrier reef fully developed. As subsidence continues the volcanic island finally disappears below sea level, and the reef becomes an atoll as in stage *C*. It is evident that in an atoll formed in this way the bed-

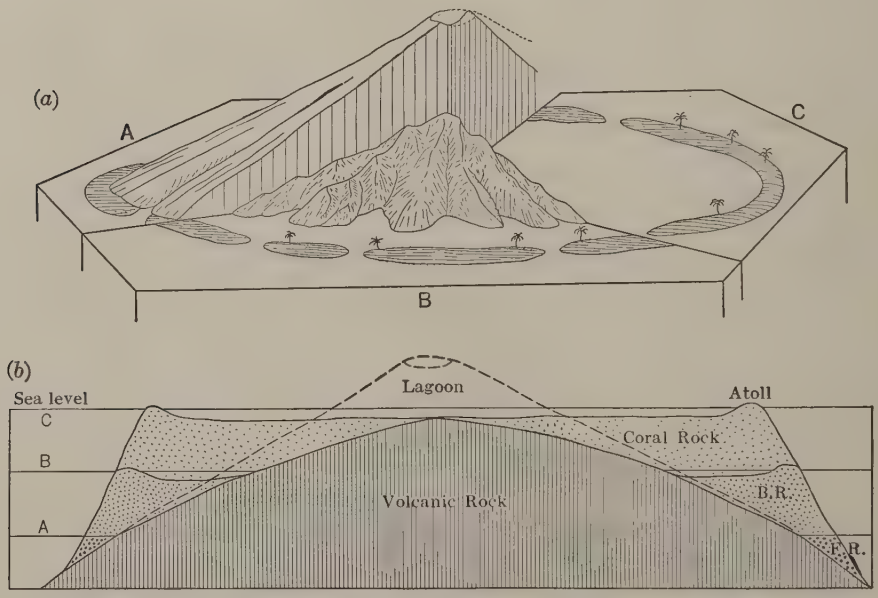

Fig. 146. — (*a*) Sector diagram of a degrading and subsiding volcanic island encircled by coral reefs. Sector *A* shows the island at an early stage of its development when its reef is fringing; *B* shows the island partly submerged and deeply eroded and surrounded by a barrier reef; *C* shows a final stage when the island has been reduced to sea level and its reef has become an atoll. (Modified from Davis.)

(*b*) Section through the island shown in (*a*) after it has become submerged and its reef has been transformed into an atoll. The successive levels of the sea correspond with the sea levels of sectors *A*, *B*, and *C*, respectively, in (*a*). Note how the fringing reef has been gradually transformed into a barrier reef and finally into an atoll through its upward growth as the island sank.

rock of the original island will have a thick veneer of coral, as in Fig. 146 (*b*). There follows as an important corollary of this hypothesis the belief that vast areas of the ocean now dotted with barrier reefs and atolls have undergone recent subsidence, in places amounting to thousands of feet.

The *solution theory*, advanced by Murray and promulgated by Alex-

ander Agassiz, sought to account for barriers and atolls without involving subsidence. It was for a time widely accepted. With a stationary sea level the reef grows outward because corals thrive best on its seaward margin. In the meantime, it is supposed by proponents of this theory, the dead portion of the inner margin of the reef, bored into by innumerable organisms, crumbles and by the combined action of solution and of scour by the tidal currents is gradually removed, while in its place a lagoon appears. Thus a fringing reef becomes a barrier. It is also supposed that some of the existing barriers began their growth around the margins of submarine platforms where islands had been partly cut away by the waves. Atolls are supposed to have grown in a similar fashion where corals grew on shallow submarine banks, as on islands that had been completely truncated by the waves. As a corollary of this theory, the coral growth in barrier and atoll reefs should be a relatively thin veneer over a rocky platform.

The *glacial control* theory, recently advanced by Daly, is in a sense a modification of the last. Daly calculates that the removal of marine water to form the continental ice caps of the last glacial period lowered the level of tropical seas from 200 to 250 feet. Owing to the colder condition of the Earth at that time (and thus of the sea), it is inferred that coral life was restricted to very narrow tropical belts. Elsewhere the oceanic islands, undefended by caps and belts of growing coral, were exposed to the erosive action of the waves, which cut wide terraces around the harder and larger islands and cut off those that were small or were formed of softer, less compact material. When the ice melted, and the seas grew warmer, the corals returned to these terraced or truncated islands, growing best at the seaward margins of the wave-cut platforms where they built up new reefs. Meanwhile the inner parts of the wave-cut platforms were drowned by the rising sea level as the glacial ice melted away, developing into lagoons. Daly sought in this way to account for the fact that most of the larger lagoons now have a depth of 200 to 250 feet and that the majority of the reef-fringed coasts show evidences of recent drowning.

In conclusion, it should be noted that these theories are not necessarily exclusive of one another. One process may have operated in some localities and another elsewhere. Nevertheless, recent chemical investigation has shown that the water in the lagoons is saturated with calcium carbonate and instead of exerting a solvent action tends to precipitate calcium carbonate. Furthermore, Davis' extensive study of the oceanic islands surrounded by barrier reefs shows that they very generally have embayed shorelines due to the drowning of their lower valleys. This feature proves clearly a relative sinking of the islands, though whether

it has been an actual subsidence of the land or a rise of sea level must be inferred from a study of the form of the barrier-encircled islands.

An island maturely dissected by streams and then slowly submerged, while protected from the waves by a growing reef, presents a shoreline marked by open embayments separated by sloping and non-cliffed spurs; but the formation of a deep wave-cut terrace at a time when coral growth had ceased must have developed a sea-cliffed shoreline, and the subsequent drowning by a rise of sea level should leave open embayments between steeply cliffed spurs. Davis (1928) has shown that while many of the islands near the margins of the tropical coral reef zone have cliffed spurs, those in the main coral reef regions ordinarily do not. The conclusion follows that glacial control has played a part in the outer areas of the tropics where reduced temperatures would first inhibit the growth of the corals, whereas in the great coral reef areas in the Pacific, subsidence must be the chief cause of the formation of barrier and atoll reefs. Recent verification of such extensive subsidence in the Marquesas Islands group has come from a study of the distribution of the flora of the islands, which indicates a recent subsidence of from 3000 to 5000 feet.

Finely divided calcareous sediments are now accumulating extensively on warm shallow submarine banks like that bordering southern Florida and the Bahama Islands where little terrigenous sediment is supplied to the sea. The conditions under which these deposits are being formed are discussed on pages 222–223.

Oozes of the Ocean Bottom. —The abyssal ocean floor, constituting three-quarters of the Earth's surface, is covered to an unknown depth by exceedingly fine-grained, soft deposits known as oozes. These sediments are of several kinds and of different sources.

Red clay is the most extensive of the oozes since it mantles the deepest parts of the ocean floor to an extent of over 50,000,000 square miles, an area about equal to the entire land surface of the Earth. Its sources are believed to be windblown dust (derived both from erosion of the land and from volcanic eruptions), pumice that has floated for a time before sinking, meteors and meteoric dust that have fallen directly into the oceans, and also the insoluble residues of organic structures that have sunk to the ocean bottom. All of this material is so thoroughly altered by chemical decay that the original source material for any sample of the clay is obscure. The red color of this extensive deposit is probably the result of the extreme slowness of its accumulation, which gives time for its thorough oxidation in spite of the deep covering of marine water.

The remaining abyssal deposits are the *organic oozes* formed largely of minute shells or shell fragments dropped by organisms that live in the

surface waters of the ocean. Some of these deposits are calcareous, being made of limy shells, and others are siliceous, being formed by the accumulation of microscopic siliceous shells.

Several distinct types of organic ooze are recognized, each being named for the group of organisms most important in its formation. The most widespread of the limy oozes is that known as *globigerina ooze* (Fig. 147). It is formed largely of the microscopic shells of single-celled animals known as Foraminifera. There are about twenty species of these animals that live floating in incredible numbers at the surface of the warmer oceans. In the reproduction of these tiny creatures the mother forsakes her shell and subdivides into a great many tiny daughters, each of which in turn secretes a new shell as it grows up, only to repeat the process. As reproduction is rapid and generation succeeds generation quickly in such forms, their abandoned shells drift down like a perpetual snowfall into the depths. Those of the genus *Globigerina* usually predominate, hence the name given to the deposits. Wherever these little shells fall below a depth of about 15,000 feet, the carbonic acid concentrated in the cold bottom waters again dissolves the shells, so that globigerina ooze is not found in the deeps of the ocean, but on the shallower parts of the abyss it mantles a vast expanse of approximately 30,000,000 square miles.

Fig. 147. — Globigerina ooze, much magnified. (After Agassiz and Murray.)

Radiolarian ooze is formed by the Radiolaria, another group of floating microscopic animals which, unlike the Foraminifera, fashion their delicate and beautifully ornate shells from silica. *Diatom ooze* is likewise formed by a group of microscopic plants, the diatoms, which secrete capsule-like shells of silica.

The oozes of the abyssal region are distinctive, and the general rarity of this type of deposit in the sedimentary rocks of the lands is believed to indicate that the continents have always remained continents and with very limited exceptions have never been covered by abyssal depths of water. Locally, rocks formed from abyssal oozes are present above sea level, as in the Island of Barbados, in some of the Netherlands East Indies, and possibly the Apennines and the southern Alps (Steinmann), but their total area probably is not great.

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CHAPTER IX

SEDIMENTARY ROCKS

Sedimentation and Stratification. — The muddy water of swollen streams is turbid with sediment in transit toward the sea. Some of this material comes to rest in lowland areas where the streams spread in flood, but the final goal for most of it is the sea floor. In either case the sediment is spread in layers of mud or sand, and much of it is compressed and cemented eventually into *sedimentary rocks.*

The layered or stratified nature of such deposits arises from an intermittent supply of sediment or from changes in the velocity of the currents. If, for example, sediments of various degrees of fineness were dropped together into still water, the heaviest and coarsest would reach bottom first and upon them would settle the next in size and so on up to the very finest. There would, of course, be no distinct layers but a complete gradation from bottom to top. But if a second lot of sediment were introduced after the first had settled, it would form a distinct layer with its coarse base resting on the fine-grained portion of the first layer. In similar fashion the sediments introduced into a lake or the sea after one storm will spread and settle before the next storm or flood. If, on the other hand, a mass of heterogeneous sediment is dropped into a steady current of moving water, the gradation takes place in a horizontal as well as a vertical direction, the successively finer material being dropped farther and farther along the bottom, or, in other words, being graded but not stratified. If the velocity of the current is altered and a second supply of sediment introduced, the new deposit will be evenly graded like the first, but its coarseness at any point will not correspond exactly with that of the first layer immediately below it. The two layers will be separated therefore by a distinct *bedding plane,* on either side of which they differ in texture. Since the velocity of streams and of the shallow marine currents changes from day to day and varies from place to place, it is inevitable that the sediments they transport will be deposited in parallel layers that differ in thickness, texture, and materials. Each single layer is known as a *stratum,* the bedded or layered deposits are said to be *stratified,* and the condition is known as *stratification* (Fig. 148).

Source and Nature of Sediments. — It has been shown in previous chapters how each of the erosive processes contributes its share to the endless stream of rock waste which the rivers move toward the lowlands and the sea. This land waste varies in coarseness from the great boulders moved by glaciers or torrential streams to the finest of mud; and it also includes material which, like salt and calcium carbonate, is borne in solution. At the same time, it varies in composition from bits of fresh and unweathered rock to the end products of chemical decay.

Fig. 148. — Regularly bedded sandstones and shales, near Pueblo, Colo. (U. S. Geol. Surv.)

It is roughly classified, chiefly on the basis of coarseness, into *gravel*, *sand*, *silt*, and *clay*.

Gravel is an aggregate of coarse sediment in which individual particles have a diameter of 2 mm. or more. The smaller pieces, ranging in size from that of grape seed to that of baseballs, are known as *pebbles*, and the larger ones as *boulders*. Besides the pebbles and boulders, gravels usually include more or less sand.

The fragments that make up the gravel are always more or less water worn. At their source they were irregular, angular pieces of rock bounded by joints or fracture planes, but in being rolled along downstream, or tossed back and forth by the waves, their angles and corners suffer the most abrasion and the pebbles and boulders become more and more perfectly rounded. Rough and subangular fragments, therefore, indicate that they have suffered little transportation and that they do not lie far from their place of origin.

In their transportation, the pebbles and boulders of softer material are the first to be destroyed by reduction to sand and gravel. Since quartz is the hardest of the common rock-forming minerals it is the predominant substance in most gravels; and well rounded pebbles and boulders, having suffered long wear, are usually composed almost entirely of quartz. On the other hand, coarse sediment that has not travelled far commonly includes other minerals and rocks, such as granite, schist, basalt, or limestone.

Sands are sediments composed of grains smaller than gravel, yet exceeding 1/16 mm. (i.e., about the thickness of this page) in diameter. Such material is coarse enough so that it will not form a coherent, plastic mass when wet. Ordinary sand is like granulated sugar in fineness. With a lens it may be seen that the coarse sand grains are rounded like pebbles by attrition, but the finest grains are usually angular. This does not mean that the larger grains have travelled farthest but rather that grains below a certain size are not rounded in transit, the explanation lying in the fact that the film of water between them serves as an effective buffer to protect such tiny particles against abrasion.

Quartz so commonly forms sand that, unless otherwise stated, quartz sand is generally understood. Many other minerals occur in sands, however, and the beaches of coral islands are in places formed of " coral sand " made wholly of calcium carbonate.

Silt and *clay* include the finest part of the land waste. This finely divided material coheres when wet, and when dry will crumble into dust. The distinction between silt and clay is largely one of fineness of subdivision, clay consisting of microscopic particles (less than 1/256 mm. in diameter) and *silt* of somewhat coarser sediment. Due to its most finely divided (colloidal) constituents, clay is plastic when wet whereas silt tends to crumble as would very fine sand. The common term *mud* is applied to either silts or clays and is used particularly where the separation of silt and clay is very imperfect.

The mineralogical composition of silts and clays is varied and much more complex than that of sand. The most characteristic constituents are hydrous silicates of aluminum and the alkaline earths and hydrous oxides of iron. Some silts, like those derived from glaciers, are produced by rock grinding instead of by chemical decay and these deposits may show a predominance of unweathered particles of all sorts of rocks and minerals.

The materials carried in solution form a large but invisible contribution to the sea, reappearing as sediments where evaporation or chemical reactions cause their precipitation, or where organisms extract the dissolved minerals to form shells.

Volcanic ash, wind-blown dust, glacial till, and organic materials such as peat are other deposited materials that enter into the formation of special types of sedimentary rocks, but although of interest and local importance, they are not of such geological significance as those mentioned above.

Places of Deposit

In Chapter IV it was explained how streams transport their burden of sediments. The spreading of marine sediment by the waves and currents has been discussed in Chapter VIII. The conditions under which the sediments finally come to rest must now claim our attention. Three major realms of deposition are generally recognized. The first is that of the land areas. The deposits formed there are known as *continental sediments* since they belong essentially to the continents, whether formed actually on the dry land or under the surface of its streams and lakes. The second realm is that of the intertidal zone and its deposits are known as *littoral sediments*. The third is that of the marine realm and its deposits, formed anywhere over the sea or ocean floors, constitute *marine sediments*. The distinction between these three realms is of such geologic significance that each of them deserves more detailed consideration.

Regions of Continental Deposits. — Although vast areas of the land surface are undergoing active erosion, about 10,000,000 square miles of the continents are now buried by accumulating sediments or by sedimentary deposits formed in the recent geologic past. These regions of deposition may be classified into four types in which the conditions of accumulation are strikingly different. These are (1) piedmont plains, (2) arid basins, (3) humid basins, and (4) great deltas.

Piedmont alluvial deposits are formed where streams debouch from relatively young and lofty mountain ranges. Here the velocity of the streams is rapidly checked as they enter the flatter piedmont belt and they drop a part or all of their load. During times of flood, when on leaving the mountains they are heavily laden with sediment, the streams spread widely beyond their channels and cover the adjacent country with layers of sand and clay. In so doing, they overflow into the lowest places, filling them with sediment, and by the continued action of this process through a long period of time they build up deposits of great thickness in the form of alluvial plains in front of the mountains. Striking examples are to be seen in the deposits that underlie the High Plains east of the Rocky Mountains, the Pampas east of the Andes, and the Indo-Gangetic Plain south of the Himalayas in India.

Since they are formed on the flanks of the mountains and well above sea level, these sediments will ultimately be destroyed as the erosion

cycle progresses, unless preserved by deep downwarping of the piedmont region. Even now, for example, the very streams that formed the High Plains are stripping these deposits away again. In this case the stripping may be hastened by climatic changes rather than by reduction of the highlands.

Desert deposits occupy great areas in the arid basins. Almost one-tenth of all the land surface of the world is embraced in interior basins, whose drainage is centripetal, with no outlet to the sea. Here the waste of the slopes constantly tends to move toward the deeper parts of the basins and accumulate in stratified deposits. At rare times of heavy rainfall the temporary streams spread widely over the lower slopes, shifting sand and gravel toward the center of the basins, while the finer material is swept on into temporary lakes (playa lakes) that form in the deeper depressions. Permanent lakes like Great Salt Lake or the Caspian Sea tend to be filled by the sediments of the larger and more permanent streams. Moreover, in times of dryness, sand is shifted by the wind. Except for wind-blown dust none of the sediment can escape and thus, through the continued action of rain wash and streams, the desert basins tend to fill with deposits. If sinking of the basins occurs while they are being silted up, the ultimate thickness of the sedimentary deposits thus formed may attain several thousands of feet. Deposits of arid basins commonly contain layers of salt and gypsum as well as thick beds of cleanly sorted dune sands. They are generally poor in fossils and exhibit light colors such as dun, buff, or pink. Some of them are red.

Deposits in humid basins also may be of vast extent. Where structural basins form rapidly in regions of abundant rainfall, great lakes are produced; but if the sinking is slow, filling may keep pace with subsidence and the region remain a swampy lowland. The second alternative is favored where the basin has bordering highlands undergoing rapid erosion, for then the incoming streams are heavily laden with sediment. A good example of this sort exists in the upper basin of the Paraguay River in South America where an area about 400 miles long and as much as 150 miles wide remains a labyrinth of lakes and swamps and channels in a low grassy plain. During the annual rainy season the whole area is flooded, but during the rest of the year only about one-fourth of it is covered. The river enters the lowland heavily burdened with sediments brought by tributaries from the Andes to the west and from the Brazilian highland to the north. The water leaves the basin fairly clear, having spread most of its sediments over the lowland to be mingled with organic matter from the decay of vegetation that flourishes abundantly in such places. This swampy condition, with resulting

accumulation of river sediment, may be indefinitely maintained if the basin is a region of continued subsidence. Though deposited in water, such sediments are to be regarded as continental in origin, since they occur in hollows of the land surfaces. Such deposits have accumulated extensively in times past and seem to represent approximately the conditions under which the older coal-bearing rocks have been formed. The abundant organic matter gives such sediments somber or dark colors.

Deposits formed in this manner are likely to escape destruction by later erosion because they lie only slightly above sea level.

A *delta* is a deposit of sediment built out by a stream into standing water. An idealized vertical section of a delta (Fig. 149) shows that it is partly above and partly below water level. Even the landward portion

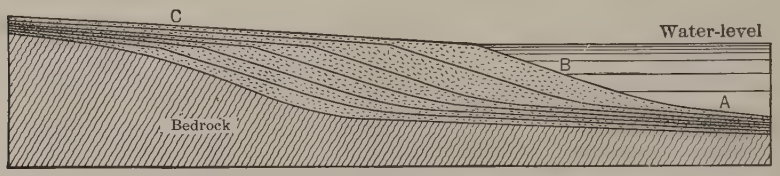

Fig. 149. — Section through a delta built into quiet water of constant level. *A*, bottomset beds; *B*, foreset beds; *C*, topset beds. (After Barrell.)

is extensively covered by fresh water when the stream is in flood and carrying its greatest load of sediment. At such times sand and silt are spread far and wide by the flood waters, forming nearly horizontal layers over the landward top of the delta. The sediment that reaches the shore is shifted by the waves and currents toward the sea, building out the shallow submarine surface like a vast embankment with a nearly horizontal surface near shore, giving way to a steeper slope beyond the line of most rapid building. The sediments that come to rest on the nearly horizontal upper surface of the delta, whether the landward or the subaqueous portion, are known as the *topset beds*. The material carried out over the inclined front of the delta constitutes the *foreset beds*, and that which is spread still farther, parallel to the sea bottom, forms the *bottomset beds*. The topset beds are partly continental but the foreset and bottomset beds are marine.

As shown in Fig. 149, when the level of sea is stationary the delta growth is chiefly seaward by an extension of the foreset beds, and in this case the continental sediments, notwithstanding their areal extent, form but a small part of the volume of the delta. When a delta is subsiding, on the contrary, more of the sediment comes to rest over its upper surface where the sluggish waters spread in flood. If the area of the landward

portion of the delta is great, it may thus happen that the volume of land deposits, formed by the topset beds, is vastly greater than that of the foreset beds which build out the delta front (Fig. 150). Consequently, a delta which is built upon a subsiding foundation tends to form dominant topset beds of continental nature. The deltas of great rivers, like those of the Mississippi and the Nile, are built out into shallow seas on the wave-swept continental platforms, and here the difference in the

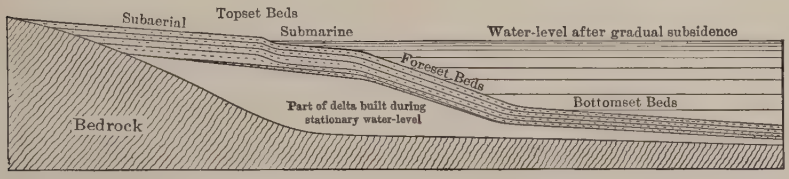

Fig. 150. — Section through a delta built into quiet water but resting on a subsiding foundation. The unshaded basal part of the delta is assumed to have formed before subsidence began. Since subsidence of the delta started, the shoreline has advanced landward and much of the sediment has come to rest as topset beds which are partly subaerial and partly submarine in origin. (After Barrell.)

inclination of foreset and bottomset beds is slight and the distinction between these two parts of the delta is not marked.

Regions of Littoral Deposits. — The total area of the present littoral zone in all the continents is probably not much over 60,000 square miles. On ordinary shores it is usually far less than 2 miles wide, though in great deltas like that of the Mississippi the salt-water marsh may be 25 miles or more in width and littoral sediments may accumulate rapidly thereon. The most remarkable extension of the existing littoral zone occurs in the so-called Runn of Cutch, a low flat on the southeast side of the Indus River delta that has an area of about 6000 square miles. It is so flat and so nearly at sea level as to be flooded with marine water during one season of the year when monsoon winds blow inland, and laid bare or covered by overflow from the Indus when the winds blow offshore. Some of the older mud-cracked marine formations may have formed under comparable conditions. In general, however, littoral deposits cannot be of great thickness, for, if the land is building out into the sea, they must give place to land deposits and be buried under them; if the sea is encroaching on the land, they must yield to marine sediments and be covered by them. In either case, they stand a rather poor chance of permanent preservation, since upon slight uplift they are exposed to stream erosion and with stationary sea level the waves gradually encroach upon the shore zone and remove them. They are likely to be preserved only in subsiding large deltas.

Regions of Marine Deposits. — The greater part of the land waste finally comes to rest on the sea floor. Most of this terrigenous (land-derived) sediment accumulates in the neritic zone but the finest muds can be transported even beyond the edge of the continental shelf and are found as much as 200 miles from land. Coarse materials accumulate in shallow water near the shore whence they are supplied, for there alone are the currents strong enough to distribute gravel and coarse sand. Fine sands extend in places to a depth of 200 or 300 feet or even more and beyond them the terrigenous muds spread as a continuous blanket not only to the continental margin but part way at least over the continental slope, where they grade imperceptibly into the oozes of the ocean floor. It is improbable, however, that much land-derived mud reaches the abyssal region. While in general there is a seaward gradation from coarse to fine sediment, it must not be supposed that all the muds are deposited in deep water, nor that there is always a shoreward phase of sandy sediment between the land and the place where mud deposition begins. In many protected embayments, as the Baltic Sea or Chesapeake Bay, muds, even of the finest sort, are now being deposited right up to the shore line.

Calcareous deposits of the neritic zone form chiefly in shallow water where clastic (fragmental) sediments are scarce or absent and where lime-secreting organisms grow luxuriantly. At the present time limy deposits are forming chiefly about the tropical coral islands and over warm shallow submarine banks such as that bordering Florida on the south and the Bahama Islands on the west. It has sometimes been wrongly inferred that marine sediments grade in depth from gravel or sand through muds to limy ooze, and it should be emphasized, therefore, that the limestones now exposed on the continents represent deposits in shallow, not deep water.

The vast deposits of red clay and organic oozes that cover the deep ocean floor have been described in the previous chapter. They are of enormous extent, but since such deposits are essentially lost to the continents and only rarely have been elevated into lands their geologic interest is limited.

Consolidation of Sediments

Stratified rocks are merely sediments consolidated into stone. Commonly the appearance of the rock clearly suggests this origin, but some of the older rocks have been modified so greatly that the resemblance to sediments would be obscure if it were not for their stratification and entombed fossils. The consolidation is the result of several factors, chief of which are *pressure* and *cementation.* As the sediments accumu-

late, the upper layers bear upon all those buried below. If the specific gravity of loose sediment is 2.3, the pressure will increase by fully 1 pound per square inch for each foot of depth. At a depth of 1000 feet, therefore, the weight of the overlying deposits is at least 1000 pounds per square inch. As a result of such compression, sediment is made more compact as the particles are squeezed tightly together and most of the water is pressed out.

Cementation results from the deposition of mineral matter in the spaces between the grains of sediment. The most common cementing substances are calcium carbonate, silica, and iron oxide. These cements may be introduced in solution in the water that fills the pore space when the sediments are deposited, or they may be introduced later by percolating ground water. Some sandstones have but little cement and therefore crumble readily into loose sand when exposed to the weather, but others have the voids almost completely filled with cement (Fig. 151) and have been thus converted into firm rock.

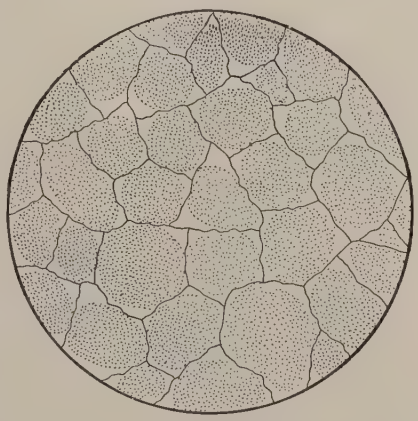

Fig. 151. — Microscopic section of a firmly cemented sandstone. The dotted areas are the rounded sand grains and the clear areas represent silica deposited between the grains, binding them into rock.

The interior heat of the Earth rising into such masses of sediments may aid in some degree to consolidate them by quickening the chemical activity of the diffused waters which deposit the cement. And finally, since the conversion of sediments into rock must be a slow process, *time* is an important element. Thus in general the more recent sediments, where they have been exposed, are softer and more friable than the older ones. It must not be inferred, however, that the process of cementation takes place only under the sea, for on land, solution by ground water, transfer to lower levels, and redeposition of cementing materials take place on a large scale.

Kinds of Sedimentary Rocks

The different kinds of sedimentary rocks depend mainly upon the nature of the sediments from which they are formed. The chief types of sediment and the equivalent sedimentary rocks are as follows:

Sediments	Compacted Strata, as Rocks
Gravel	Conglomerate
Sand	Sandstone
Silt and clay (mud)	Shale
Calcareous mud and sand	Limestone

It must not be imagined that the different kinds of rocks mentioned above are always sharply defined from one another as wholly distinct types. Just as muds grade through sand into gravel, and pure calcareous deposits into silts or clays, so may the various rocks formed from them grade into one another. In the description of these impure or mixed types of rocks, those that were muddy or clayey sediments are termed *argillaceous* (from the Greek *argillos*, clay). Similarly, sandy ones are described as *arenaceous* (from the Latin *arena*, sand) and limy sediments or rocks are described as *calcareous* (from the Latin *calx*, lime). These three adjectives are used in the discussion either of sediments or of the rocks derived from them.

The chief types of sedimentary rocks deserve more detailed discussion.

Conglomerate. — A typical conglomerate is shown in Fig. 152. It consists of rounded pebbles set in a matrix of well-cemented sand. Generally the pebbles and boulders in conglomerates are made of quartz, but they may consist of fragments of any kind of rock. The fragments vary in size from a small fraction of an inch to many feet in diameter. They may be packed together with little fine matrix or mingled with any proportion of fine material. If the pebbles are few and scattered through a sandy matrix, however, the rock should be called a *pebbled sandstone* rather than a conglomerate.

The pebbles and boulders may be well rounded or, if they have suffered but little transportation, they may be more or less angular. If they are distinctly angular the rock is a variety of conglomerate known as *breccia*.

A special type of conglomerate or breccia is produced when the layers of accumulating sediments, indurated before complete burial, are broken up and the dislocated fragments rolled about and then recemented. This happens where exceptional storm waves violently stir up the bottom, where layers exposed to the atmosphere are broken up by mud cracks, where streams in their floods undercut the sides of their channels, or where mountain-making disturbances rupture the beds on the sea floor. The conglomerates thus formed represent merely an episode in the deposition of the formation in which they lie and they are therefore known as *intraformational conglomerates*. They are made of

the same materials as the enclosing beds and do not signify uplift and renewed erosion as do the other types of conglomerates formed of imported gravel. Usually the individual fragments of the intraformational conglomerate are only imperfectly rounded and they may be quite angular.

Arkose is a special variety of conglomerate or sandstone containing much feldspar. Its occurrence indicates that the component material was not long exposed to weathering before it was deposited, and therefore probably was not transported great distances. Arkose is more com-

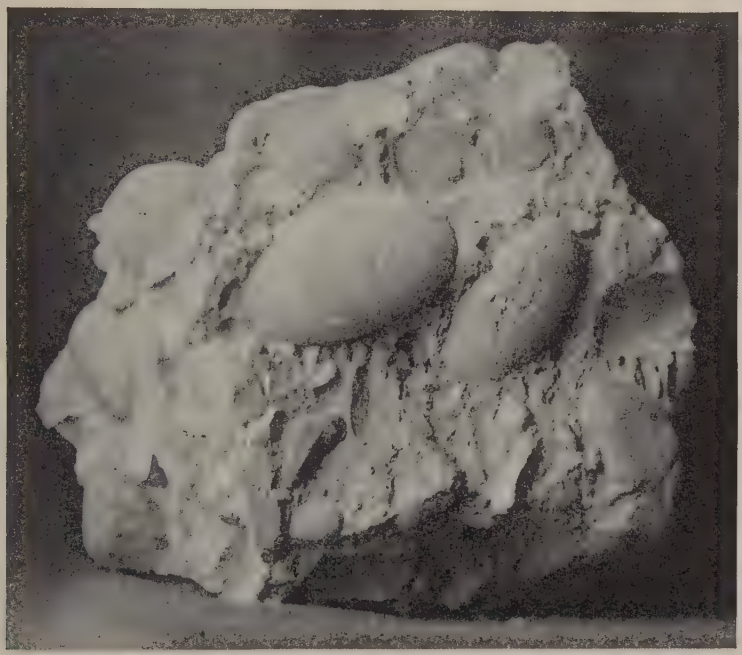

Fig. 152. — A piece of conglomerate, shown about half natural size.

monly of continental than marine origin, and its formation is favored by the breaking down of granitic rocks in cold or arid climates where rock decay is inhibited.

Sandstone. — Sandstones are composed of cemented sand grains, which consist most commonly of quartz. Many sandstones are quite even in grain, but there are hybrid varieties grading on the one hand into conglomerate, and on the other into shales. In red and brown sandstones the cement is mainly oxide of iron but in white, buff, or gray varieties it is most commonly either silica or calcium carbonate. Sandstones are generally porous and the interspaces may amount to 30 per

cent of the total volume. For this reason they are favorable reservoirs for artesian water and petroleum.

Shale. — The indurated equivalent of silts and clays is shale. The majority of shales possess a more or less thinly laminated structure resulting from closely spaced, parallel bedding planes. Such shales tend to split easily along these natural planes of stratification. If the laminae are thin and flexible the shale is said to be fissile. On the other hand, some shales occur in layers of considerable thickness that show no subdivision by bedding planes and tend to break into blocks instead of laminae. Such massive or blocky shales are sometimes designated *mudstones.*

Shales are soft and generally weak rocks, crumbling readily into small chips. They show a great variety of colors as do the muds from which they form. Unlike sandstone, they tend to be impermeable to water. Since they are composed to a greater or less degree of clay they yield a strong and characteristic clay odor.

Shales occur in both continental and marine deposits. In the former, they represent mostly the flood-plain deposits well back from the channel where roily waters, impounded as backwater or at least retarded in their flow, have been unable to keep the fine sediments in suspension.

Black shales constitute a distinctive type of sedimentary rock commonly associated with coal beds and also occurring abundantly in marine deposits, where they are important as source rocks in the geological formation of petroleum. The dark color of these shales is due to carbon derived from the incomplete destruction of organic matter buried with the mud. An abundant supply of organic matter and stagnation of the water appear to be the essential requirements for black shale formation, the stagnation leading to a deficient supply of dissolved oxygen so that decay is inhibited. Depth is not a vital factor, though obviously any sharply depressed areas or hollows in the sea floor are likely to be passed over by bottom currents and to become stagnant black mud holes. Black muds are now accumulating in both shallow and deep water. Where black muds are forming, hydrogen sulphide (H_2S) is usually generated and it in turn reacts with iron salts in the sea water to form pyrite (FeS_2), which is precipitated in the form of concretions in the black shale, or very commonly in the form of replacements of fossil structures.

Limestone. — Limestone is the consolidated equivalent of limy ooze, calcareous sand, or shell fragments. It is composed of the mineral calcite or calcium carbonate ($CaCO_3$). Pure varieties of limestone are white or light gray, but impure varieties are buff, brown, red, dark gray, or even black. If fine-grained and nearly pure, the rock is compact

and tough and forms one of the strongest of structural stones; but with an admixture of clay, limestones grade into calcareous shale which weathers readily into clay. Some limestones have a dense texture, others are finely crystalline, and still others distinctly granular. Some impure varieties appear earthy.

Chalk is a soft, porous variety of limestone. Some chalks are composed largely of the microscopic shells of Foraminifera, the tiny animals that also make the globigerina ooze of the ocean floor. These minute shells are extremely fragile and commonly occur much broken. In addition to the shell fragments there is normally a matrix of fine particles of calcium carbonate that has been considered to be a chemical precipitate. Some chalky deposits, as those of Kansas and Alabama, are made up largely of such finely divided material, with an admixture of clay, and show but few of the foraminiferal shells.

Chalk was once supposed to represent deep-water deposits like the modern oceanic oozes; but in spite of the fact that globigerina ooze will probably form chalky deposits, it is now known that the great chalk beds of the present lands include shells of shallow water animals and are associated with coarse-grained sediments that could not have been washed into deep water.

Coquina is a variety of limestone made up of shells and coarse shell fragments heaped together and loosely cemented.

Dolomite is a rock formed largely of the magnesium-calcium carbonate mineral, dolomite, $CaMg(CO_3)_2$. It may be considered a special variety of limestone. In fact, there are probably all gradations between calcite and dolomite limestones, and the mixed types are more common than pure dolomite. Dolomite resembles calcite limestone in most of its characters and is commonly regarded as limestone by others than specialists. It is of great extent and importance in the older rocks, especially. The Dolomite Alps of the eastern Tyrol, for example, have been carved out of dolomite formations 3000 to 4000 feet thick, and equally great masses of this kind of rock occur in Alabama and in western Texas.

Limestone represents the calcium carbonate that is carried in solution to the sea and there precipitated; but the conditions for its deposition are less obvious than those under which the clastic (fragmental) sediments are formed and there is still much to learn about the relative importance of the different agents that precipitate the limy sediments.

The soluble bicarbonate of calcium is very unstable and can easily be made to give up one molecule of its CO_2 according to the formula $H_2Ca(CO_3)_2 = CO_2 + H_2O + CaCO_3$; but when this happens the corresponding molecule of $CaCO_3$ becomes insoluble and is precipitated.

Recent investigations have shown that the warm shallow marine water is essentially saturated with calcium carbonate whereas the colder, deeper water is undersaturated because of its richness in CO_2. Wherever the deeper water rises in ocean currents to flow over shallow submarine banks in tropical regions the rise in its temperature tends to drive off part of the CO_2 and to leave the warmed water supersaturated with $CaCO_3$. Chemical precipitation would occur in such places if the lime were not rapidly extracted by organisms before the saturation point is attained. However, these places afford the most favorable conditions for animals and plants that are prodigal in their use of carbonates to form

Fig. 153. — Mudflats off the west coast of Andros Island at low tide, showing the nature of the limy deposits that cover large areas of the Bahama Banks. The sediment is a soft white paste of nearly pure calcium carbonate. (R. M. Field.)

shells or skeletal structures. Corals thrive and form reefs on such tropical shoals and with the corals are associated lime-secreting algae and a host of shell-forming creatures.

Impressed by the limy mud and coral sand formed about coral reefs, the earlier naturalists attributed to corals a predominant rôle in the precipitation of the limy deposits. Many limestones, however, present little evidence of a coralline origin and recent investigations have shown that modern calcareous sediments of great extent are being precipitated directly as a fine ooze or mud. One of the most extensive areas of modern limy sediment is that of the Bahama Banks south of the Florida Straits. The total area of this great shoal is over 7000 square miles and the average depth of its water is less than 20 feet. The bank is formed

of limestone and great areas of its surface are mantled by fine white limy ooze (Fig. 153). The agent of deposition of this limy material is still uncertain. Coral reefs are limited to the extreme margins of the shoal and appear not to be large contributors to the deposit at the present time. It has been thought that a bacterium (*Pseudomonas calcis*) caused the precipitation of the calcium carbonate but doubt has recently been cast on this idea. The relative importance of chemical and biological agents in the deposition of the carbonate sediments is therefore still unknown.

Diagenesis of the Calcareous Sediments. — Important chemical alterations not uncommonly take place in the limy sediments during or shortly after their deposition. For example, no dolomite is formed directly, either by chemical precipitation or by organic secretion. The deposit goes down first as calcite and then, while still in contact with marine water, some of the calcite is replaced by magnesium.

Another type of alteration is to be seen in the solution and recrystallization of the fine particles of limy mud to form granular or coarsely crystalline limestones, or, about coral reefs, to transform the loose coral débris into compact stone.

All such chemical alterations as modify the sediments while they are accumulating are embraced under the term *diagenesis* (Gr. *dia*, in two parts, + *genesis*, birth).

Minor Sedimentary Rocks. — *Coal*, *iron ore*, *rock salt*, *gypsum*, and *chert* or *flint* are sedimentary rocks of special interest, but from the geologic point of view they occur in volumes so limited in comparison with the enormous bulk of the shales, sandstones, and limestones that they are of little importance considered merely as rock masses.

Coal and iron ore are given special consideration in Part II of this book.

Characteristic Features of Sedimentary Rocks

In addition to the stratification which these rocks always exhibit, other features are commonly displayed which throw light upon the geologic history of the rocks. Among these features are fossils, mud cracks, ripple marks, and cross-bedding.

Fossils. — Fossils (Latin *fossiles*, from *fodere*, to dig up) are the remains or imprints of animals or plants that were buried with the accumulating sediments (Fig. 154). The organisms are rarely preserved entire; generally only the hard structures, such as shells and bones, endure. These parts may be preserved without change, but commonly the original material is replaced, one molecule at a time, by mineral matter so that the fossil becomes a solid stony object or *petrifaction.*

Not uncommonly the organic structure is wholly removed, leaving only an imprint or hollow mold in the rock. *Natural molds* of this sort are not infrequently filled by subsequent mineral deposits which then form *natural casts* of the organic objects. Finally, footprints or trails made by animals crossing soft mud are preserved under favorable conditions.

Since most fossils represent the animals and plants that were living where the sediments accumulated, they throw much light on the condi-

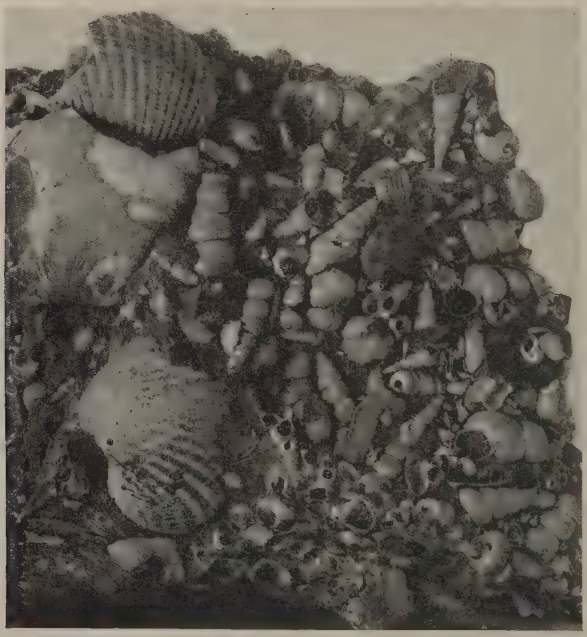

Fig. 154. — Fossil shells in stone.

tions that prevailed during the deposition of the fossiliferous rocks. They show, for example, whether the beds were laid down on the land or beneath the sea, since continental sediments contain the remains of land plants, bones of land animals, or shells of river clams, whereas marine sediments contain representative marine shells. Something of the climate of the region at the time of deposition may be indicated also; for example, fossils of tropical plants and animals, such as palms and alligators, are found associated in certain formations of the "Badlands" of South Dakota. The manner in which fossils are used in deciphering the past history of the Earth and its inhabitants is considered in detail in the second part of this book.

Mud Cracks. — Soft muddy sediments left exposed after the recession of flood waters shrink and crack into characteristic polygonal blocks like those shown in Fig. 155. These desiccation fractures are known as *mud cracks*. Further exposure to air and sun bakes and hardens the blocks of mud. During the dry season wind-blown sand or silt may cover the surface, filling the cracks with sediment coarser than the mud-cracked layer. In this way the form of the polygonal blocks is preserved

Fig. 155. — Modern mud cracks formed on the delta of the Colorado River. (U. S. Geol. Surv.)

even after the return of later flood waters. After the whole series of deposits has subsided and become hardened into rock, the layers of shale and sandstone may later be exposed, exhibiting these "fossil" mud cracks on the bedding planes. When the beds are thus exposed, the softer, mud-cracked layers commonly crumble away, leaving natural casts of the mud cracks as projecting ridges on the lower surfaces of the sandstone layers. Obviously, the conditions that favor the formation and preservation of mud cracks are also ideal for the preservation of footprints of animals that have crossed the soft mud soon after its deposition, and therefore mud cracks and footprints are frequently associated.

The most favorable places for the occurrence of mud cracks are on the flood plains of large rivers, the landward portions of great deltas, and the wide flat shores of shallow interior lakes that shrink or disappear during the dry season. In spite of its alternate wetting and drying, the littoral zone is unfavorable for the formation of mud cracks because the mud has not sufficient time to dry out thoroughly before the return of the tide. Mud cracks may, however, form to a limited extent at the upper margins of estuaries where the spring tides reach only for a few days in each month. The mud cracks can form only when the sediment is exposed to the air and allowed to dry and shrink. They are generally lacking therefore in marine deposits.

In the early Paleozoic formations, however, there are well-known examples of mud-cracked marine limestones. They represent very

Fig. 156. — Current-formed ripple marks exposed at low tide near Windsor, Nova Scotia. The movement of the current was from left to right. (Geol. Surv. of Canada.)

special conditions that may find an analogy at present in the Runn of Cutch or in the Bahama Islands. During the hurricanes of the spring of 1928 the lower islands and the borders of the larger islands of the Bahama group were flooded by the seas that swept over the Bahama Banks (see page 222). It was later observed that many square miles of this sediment were mud-cracked. In places limy sand and fragments of marine shells have since blown over it. It is quite possible that in the past similar conditions have permitted the covering of extensive low coastal lands by marine sediments laid down during exceptional hurricanes beyond the normal confines of the sea.

Ripple Marks. — Where currents sweep granular sediments along the bottom, the surface of the deposit develops parallel ridges resembling the ripples on the surface of a pool of water. These are known as *ripple marks.* They are also formed on the land where sand is shifted by the wind, as on sand dunes, or under running water. Ripples thus produced retain their form and migrate slowly with the current, because the sand grains are rolled up the windward or stoss side and fall down the leeward slope. Current-formed ripples, therefore, have an unsymmetrical form, the stoss side being a gentle slope and the opposite side steeper. The direction of flow of the current is thus autographed in the form of the ripple marks (Fig. 156).

Where oscillatory waves touch bottom they also develop ripples as a result of the to-and-fro motion of the bottom particles; but oscillatory ripples, in contrast to current-formed ripples, are symmetrical (Fig. 157).

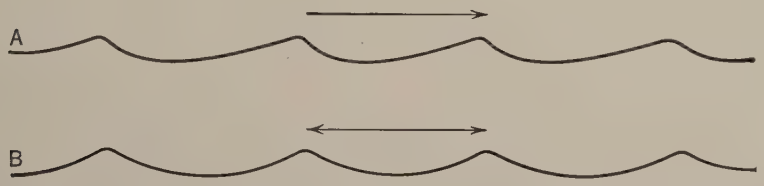

Fig. 157. — Unsymmetrical profile of current ripples (*A*) contrasted with the symmetrical profile of oscillation ripples (*B*).

Current ripples may be formed wherever currents disturb sandy or silty surfaces, whether it be wind currents on the land, the currents of running streams, or any of the currents in the seas. Waves of oscillation, on the other hand, occur only under standing water and in depths touched by wave action. Ordinary storm waves in the sea are ineffective below 200 or 300 feet, but exceptionally the bottom is rippled to depths of 600 feet or more.

Cross-bedding. — In many deposits of coarse detritus, such as conglomerate and sandstone, the layers of particular beds are inclined to the general planes of stratification at considerable angles (Fig. 158). This structure is known as *cross-bedding.* It is produced where sand is shifted by either wind or water currents in such a way as to be spilled down the front of an advancing deposit; as the foreset edge of a delta, the front of a gravel bar in a stream, the front of a sand dune, or merely the front of a current ripple. In any case the individual layers or laminae come to rest on a slope inclined to the general surface of deposition.

The scale of the cross-bedding may be great or small. In dune sands single cross-bedded layers are commonly tens of feet thick but in silts

shifted by small ripples the cross-bedded layers are fractions of an inch thick. Since the inclination or foresetting of the laminae is always downstream, it is possible to infer the direction of the currents that moved the cross-bedded deposits.

The cross-bedding produced in dune sands is one of the most distinctive types (Fig. 158). The laminae are inclined first in one direction and then in another as a result of the repeated change of direction of the winds. Moreover, the front slope of the dune is not straight but curved where the sand rolls forward at its base as a result of eddies in the wind

Fig. 158. — Cross-bedding of the type characteristic of dune sands. Navajo sandstone near Glen Canyon, Utah. (U. S. Geol. Surv.)

on the leeward side of the dune. As a result, the cross laminae, developed on a large scale, descend in curves that become tangential to the bedding planes below.

Concretions. — Stratified rocks in many places contain inclusions called concretions. These objects differ in composition from the enclosing rock and are generally rounded or nodular in form; some are quite spherical, others flattened, ovate, elongated, ring-shaped, or compound; still others exhibit odd and fantastic shapes. They range in size from a fraction of an inch to many feet in diameter. Ordinarily they are formed from one of the minor constituents of the rock; thus, in chalk and limestone they are composed of silica; in sandstone, of iron oxide or carbonate of lime; in shale, of calcium carbonate or sulphide of iron. Although some are pure they commonly contain large amounts of the inclosing rock material, and the planes of stratification of the rock can be seen passing through some concretions.

Their origin appears to lie in the solution of some of the minor constituents in the rock and the redeposition of this material around certain centers as nuclei. Very commonly they contain at their center a fossil and in such cases it appears that the products of decay of the buried organism have caused the precipitation of the mineral matter, the organism serving as a nucleus. Remarkable imprints of fern leaves, insects, and marine animals are obtained by splitting such concretions. The shells and bones of even large animals are found in some concretions.

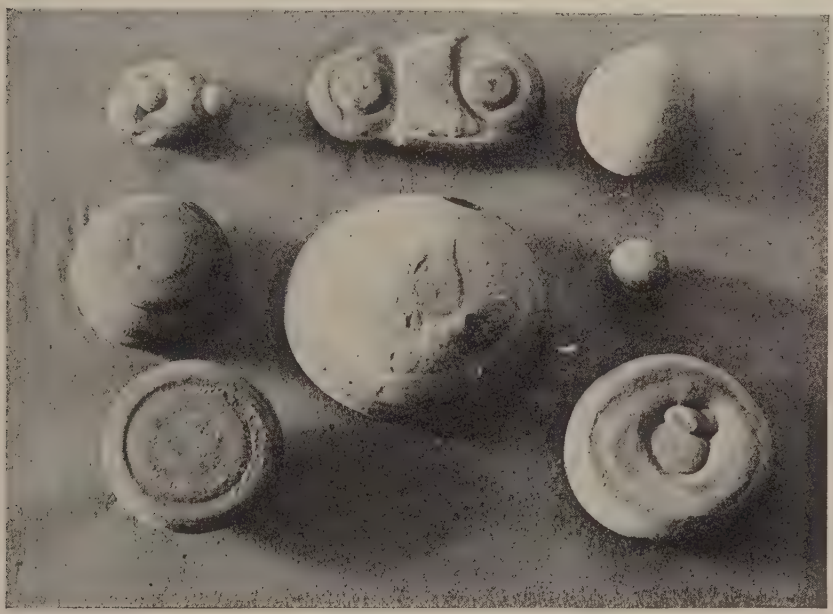

Fig. 159. — Concretions from clay beds, Long Island.

Some concretions appear to have formed about nuclei of inorganic substance, as grains of sand, and still others show no definite nucleus of any sort. Iron oxide concretions are not uncommonly hollow or have only a partial filling of loose sand.

Colors of Sedimentary Rocks

Sedimentary rocks present a great variety of colors, some of which help to indicate the conditions under which the sediments accumulated. Sandstones are usually light-gray, greenish-gray, buff, brown or red; shales generally light-gray, dark-gray, greenish-blue, red, purple, maroon, or black; limestones are nearly white if pure but gray, buff, brown,

pink to red, or black if impure. Some of the more recent deposits of clay are nearly white or have delicate shades of pink or lavender. The coloring matter is usually carbon or oxides of iron.

Black or dark-gray sediments owe their color essentially to the black carbon resulting from the partial destruction of organic matter that was buried with the sediment. The black shales are especially rich in organic matter, and coal represents concentrated deposits of this nature.

Oxides of iron are the great coloring agents in nature. Iron is the fourth most abundant element in the Earth's crust and iron-bearing minerals are disseminated throughout almost all rock masses. There are few sediments, therefore, that are completely wanting in iron. In the igneous rocks the iron occurs generally in the ferrous state combined in the silicate minerals and in this condition is not strongly colored. In weathering, however, the iron becomes oxidized and then takes on bright hues of yellow, brown, or red like those displayed by rusting iron. Iron oxide has the capacity of combining with a variable proportion of water to form hydrated iron oxides and the shade of color from yellow through brown to red seems to depend largely on the proportion of adsorbed water. The dehydrated ferric oxide (Fe_2O_3) is of deep red color.

Red sediments are commonly associated with beds of rock salt and gypsum that have been precipitated through the evaporation of large bodies of salt water. They also commonly make up or accompany sandstones and conglomerates containing unweathered feldspars. This common association of red sediments with the phenomena of arid regions has inclined many to regard all red formations as the products of arid climate. Against this belief there is, however, the anomalous fact that the great areas of modern red soil are in the warm and humid regions, and most of the deserts have dun or brownish soils. Since the sediments transported by streams are all recruited from regolith, it appears, therefore, that the greatest sources for red sediments are warm and somewhat humid lands. However, red soils form where there is sufficient relief to insure free circulation of the oxygenated water through the regolith; and in the lower, swampy regions, vegetation is luxuriant, the stagnant waters are quickly robbed of their oxygen and here strongly reducing conditions exist. Therefore, even if red oxides be introduced into swampy, humid regions they tend to disappear and the sediments turn dark because of the excess carbon. Since sediments are generally deposited in the lowest places, it follows that in spite of their original color they have small chance of forming red deposits in humid basins. But life is sparse in the desert basins and the soil is dried and oxidized to great depths between the infrequent rains. No reducing conditions

obtain here and if red sediments are washed in from the more humid uplands or slopes they are likely to remain red in color.

Due to the abundance of marine life, reducing conditions generally obtain on the sea floors and, consequently, marine sediments are not commonly red. Nevertheless where red sediments are swept into the margin of the sea in sufficient quantity there may not be enough organic matter on the bottom fully to reduce the iron oxides. Off the mouth of the Amazon, for example, there is an area of modern red muds, and a number of cases could be cited among the sedimentary rocks where abundant marine fossils are entombed in red strata. The color of the abyssal red clays is apparently due to the sparseness of life on the ocean bottom and to the extreme slowness of accumulation which permits thorough oxidation of the fine sediments.

In conclusion it should be evident that red strata occur chiefly in the continental formations and especially those that accumulated in arid or semiarid environments, but that red color is not of itself proof of any one condition of deposition.

Stratigraphic Relations

Much light is thrown on the geologic conditions under which a group of strata was formed by a study of the distribution of the beds, their relation to other rock masses, and, especially, the distribution of the fine and coarse sediments in the group. Some of the more important of these stratigraphic relations are discussed in the following paragraphs.

Relative Age of Beds. — In view of the origin of stratified rocks as superposed layers of sediment, it is evident that each layer is younger than the next below. Except where the rocks have been overturned, or broken and thrust out of their normal sequence as in some mountain zones, the youngest is at the top and the oldest at the bottom of any group of strata. Many conclusions as to the historical sequence of geological events rest upon this evident and fundamental law of stratigraphy.

Grouping of Strata into Formations. — A group of similar strata, closely related in their development, constitutes a geological *formation*. The view of the Grand Canyon wall seen in Fig. 160, for example, shows six great formations. The lowest includes intergrading beds of sandy shale and shaly sandstone more than 1000 feet thick; the next is a rather pure and homogeneous limestone that outcrops in 500-foot cliffs along the middle of the Canyon wall; the third is a thick group of shaly and sandy beds; the fourth a homogeneous shale; the fifth a pure, cliff-making sandstone; and the uppermost a limestone. During

the time of the formation of the lowest group of strata, conditions of deposition must have remained nearly uniform and the processes supplying fine sand and mud to this region presumably operated with more or less continuity. A different set of conditions then came into existence and endured while the thick limy sediments of the second formation were produced. Eventually conditions changed again and yet again

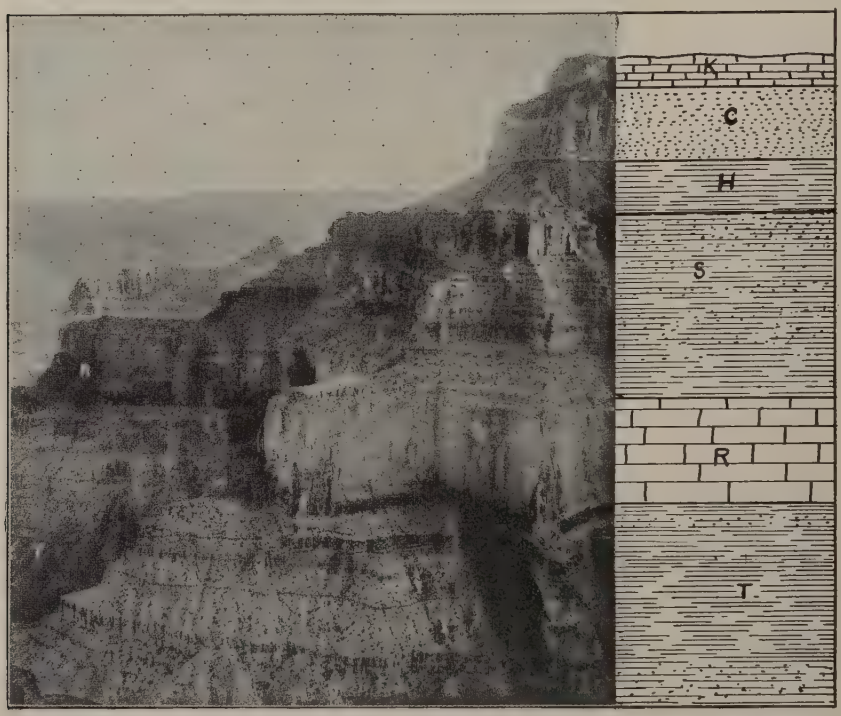

Fig. 160. — The south wall of the Grand Canyon of the Colorado River near El Tovar, showing six great geological formations, each of which is marked with a letter in the explanatory diagram. (U. S. Geol. Surv., and Ariz. Bureau of Mines.)

to give rise to the higher groups of strata. Thus each of the formations is the product of a distinct set of formative processes.

Ideally a formation is a homogeneous group of strata, as limestone or sandstone or shale, but it may also be a group of interbedded layers of different kinds of rock, as the shale and sandstone in Fig. 148. All sedimentary rocks are divided by geologists into formations, but the formational units are not generally so ideally simple and homogeneous as those shown in Fig. 160. For example, the 200 feet of strata exposed in the river bluffs about Kansas City are all so closely related that they are embraced in a single formation, although they include a series of alter-

nating shale and limestone beds. Minor groups or subdivisions of such a formation are distinguished as *members*. Figure 161 shows five members of the Kansas City formation. Not uncommonly it is difficult to decide upon the limits of such formations and the grouping is more or less arbitrary.

It is the universal custom to give rock formations geographical names. Thus, certain limestone beds well exposed about St. Louis are known as the *St. Louis formation* and the name is extended to them wherever they can be traced or identified. Likewise the beds forming the river bluffs

Fig. 161. — A portion of the Kansas City formation at Kansas City, showing five distinct *members*, two of which are of shale and three of which are of limestone. (Dunbar.)

about Kansas City and extending far away north, east, and south, constitute the *Kansas City formation.*

Intergradation of Different Kinds of Sedimentary Rocks. — Sediments vary from place to place in their coarseness and in their composition. Gravel and coarse sand accumulate where transporting currents are able to carry them, but grade laterally into finer sand where the currents slacken. Fine sand in turn grades into mud in the sheltered places and in quiet depths of water. The limy deposits of clear shallow sea bottoms grade eventually into muddy or sandy deposits near degrading lands. The change from one type of sediment to the next is always gradational and seldom abrupt. It is inevitable, therefore, that sedimentary rocks of one kind grade into contemporaneous strata of different composition. In general, coarse sediments grade most rapidly into other

types of deposit and are most localized and irregular in their distribution. Thus conglomerates, which mark old stream channels or shore zones, generally form irregular linear deposits extended parallel to the stream course or the shoreline but in other directions grading into sandstone within a few miles at the most. Coarse sandstones generally show a similar inconstancy but single beds of shale or limestone may cover thousands of square miles.

In a basin receiving sediments for a considerable time, gradations will occur not only in a lateral sense, but from bottom to top of the deposit, as the erosion cycle progresses. If, for example, a land area were uplifted and then slowly degraded the sediments derived from it and deposited in adjacent depressions would show a regular cycle of deposition (Fig. 162), with coarse material near the base succeeded by finer and finer sediment. The sedimentary cycle is a corollary of the erosion cycle. In the youthful and submature stages of dissection the streams can transport abundant coarse sediment but, as the land is worn lower, the streams flow more slowly and carry only fine material, and finally, late in the cycle of erosion when mechanical sediments are reduced to a minimum, calcareous deposits may form near the shore. If the land area is then rejuvenated, rapid erosion begins again and a new cycle of deposition is inaugurated, clastic sediments coming to rest upon the limestone.

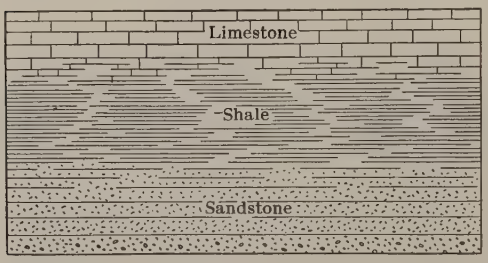

Fig. 162. — To indicate the normal succession of sedimentary deposits derived from a degrading land mass.

Actually, the sedimentary cycle is seldom if ever brought to completion in the simplicity pictured above, for it may be interrupted by many factors as, for example, uplift of the land, increased erosion due to climatic changes, uplift of the region of deposition or change in the course of marine currents.

Overlap. — Where the sea (or the area of deposition) gradually encroaches upon a land surface, the beds have the relations shown in Fig. 163, each layer overlapping the next below and extending some distance beyond it so as to thin out against the old land surface. This relation, known as *overlap*, is of common occurrence and great geologic importance. It should be noted that the sands, being near-shore deposits, grade laterally into muds and that they are not of one age. At any place the sand is succeeded by mud but in successive sections

from right to left the sand rises higher in the sequence as it follows the encroaching shore line.

If, on the contrary, sea level should remain constant and accumulating sediments extend the land surface seaward, each bed may fall short of

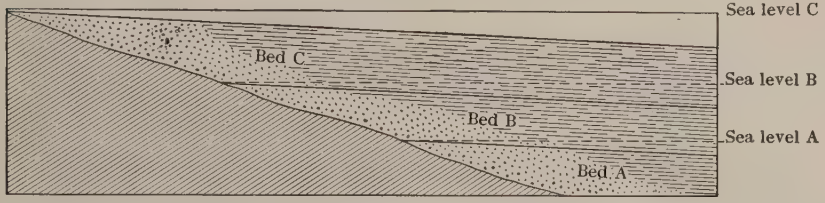

Fig. 163. — To illustrate the progressive *overlap* of beds *A*, *B*, and *C* as a result of the rise of sea level by three corresponding stages. Note that the shoreline has moved progressively to the left. Symbols for sandstone and shale same as in Fig. 162.

the next below as in Fig. 164, producing the relation known as *offlap*. It should be noted that in this case coarser sediment is carried out over finer as the sea floor is silted up. A gradual lowering of the sea level will also produce offlap as the accumulating sediments are then shifted farther seaward at successive stages of the retreat of the shore.

Alternations of subsidence and building lead to an interfingering or dovetailing of finer and coarser sediments. The same phenomenon may

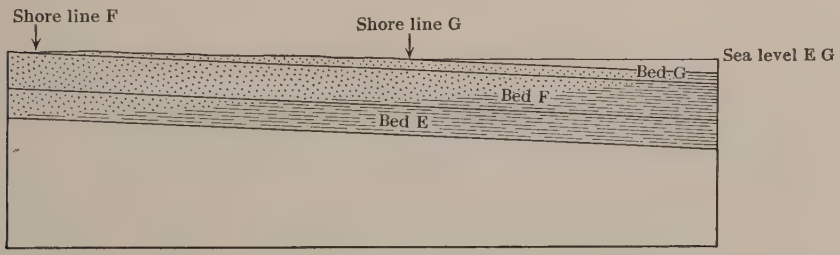

Fig. 164 — To show *offlap* of beds *E*, *F*, and *G* as a result of seaward building of the sediments while the sea level remained constant. Note that the shoreline has moved progressively to the right. A gradual lowering of sea level would produce similar results. Symbols for sandstone and shale same as in preceding figures.

be produced by fluctuations in the rate of erosion or in the strength of currents. For example, during times of great floods or of stormy seas, coarse sediment is distributed farther than in times of quiet and may spread as layers of sand or silt over mud bottoms. Likewise the mud stirred up by exceptional storms may spread into regions of limy deposits. The interbedding of thin layers of limestone and shale or of sandstone and shale, like that shown in Fig. 148, and in the shaly formations of Fig. 160, illustrates the results of such fluctuations.

Extent and Form of Sedimentary Formations. — In view of the discontinuous nature of the regions of deposition it is evident that no geologic formation can be world-wide in its extent. Beds of mechanical sediment, as sandstone and shale, imply land surfaces from which they were derived, and basins in which they were laid down; obviously their areal extent must be limited by the borders of the basins next to which the sediments thin and disappear. In geometrical form a group of sediments is relatively broad and sheet-like, consisting of subparallel layers or strata. If deposited in a circumscribed basin the deposits will be roughly lenticular and thickest in the deepest (or most rapidly subsiding) part of the basin; but if laid down along an open sea coast the form of the deposits tends to be wedge-like, thickest near shore and progressively thinner toward the ocean.

Unconformity. — One of the most important stratigraphic relations is that of unconformity, a subject that is treated in Chapter XII.

READING REFERENCES

Treatise on Sedimentation; by W. H. Twenhofel. 661 pages. Williams and Wilkins Co., Baltimore, 1926.

A source book for the study of both sediments and sedimentary rocks, with many references to special literature.

CHAPTER X

VOLCANOES AND VOLCANISM

The various agencies that modify the surface of the Earth, such as the atmosphere, the water in its several forms as rivers, seas, and ice, and plant and animal life, derive the energy that enables them to do their work from a source exterior to the Earth: from the sun. For without the sun these activities would cease and the Earth's surface would be inert. Toward these agencies the Earth is passive, except as it adds the force of gravity to help them in their work.

We have now to consider a set of agencies that also are modifying the Earth's surface, but whose energy is derived from sources within the Earth itself. So far as we can judge they are due either directly to the interior heat of the Earth or to changes going on within the Earth that produce heat. We shall describe first the results that they accomplish at the surface, and then inquire into their origin.

VOLCANOES

General Description. — A volcano is a hill or mountain composed of materials amassed around a vent through which they have been ejected from the Earth's interior in a highly heated or molten condition. The typical volcano is conceived of as a steep conical mountain with a pit-like crater at its top, from which issue from time to time gases, ashes, bombs, and flows of molten rock called *lava*. The ejection of material is termed an *eruption*, and to the human mind volcanic eruptions are perhaps the most impressive of all geological phenomena, from the immensity of the forces displayed, the magnitude of the results achieved, and the disastrous consequences that they frequently entail. Volcanoes vary widely from the typical form: some are low and of gentle slopes, or high and steep; conical, or elongated and irregular in shape; and the crater may be at the top, or on the side, and it may be of variable shape, or even be wanting.

In size volcanoes range from small cones 100 or 200 feet high to some of the loftiest mountains on the globe. Thus certain of the highest peaks of the Andes are volcanoes; some of these are still active, as Cotopaxi in Ecuador, 19,600 feet high, with a crater half a mile in diameter and 1500 feet deep, whereas others, like Aconcagua (23,000

Fig. 165. — Mt. Hood, Oregon. A composite volcanic cone, somewhat modified by glacial erosion. (A. M. Prentiss.)

feet) and Tupungato (21,500 feet) on the border between Chile and Argentina, and Chimborazo (20,500), in Ecuador, which apparently have no craters and are not now active, have become extinct in the recent geologic past. These volcanoes are built upon a dissected platform of much older rocks, above which they rise 10,000 to 12,000 feet; but the volcanoes of the Hawaiian Islands rise from the bottom of the Pacific Ocean at depths of 14,000 to 18,000 feet, and their highest summits project 14,000 feet above sea level, thus making the total height 30,000 feet. In the United States the higher peaks of the Cascade Range, beginning on the north with Mt. Baker (10,703 feet), Mt. Rainier (14,408), and Mt. Adams (12,307) in Washington; Mt. Hood (11,225) Fig. 165, in Oregon; and Mt. Shasta (14,162) in northern California, are volcanoes, which are now dormant or have recently become extinct. Lassen Peak (10,460 feet), standing at the south end of the Cascade Range, is the only active volcano within the continental United States. Mt. Etna, on the coast of Sicily, rises 11,000 feet above sea level, and the diameter of the base of its cone is 30 miles. Its lower slopes are gentle and studded with many small, or " parasitic," cones.

Character of Eruptions. — Three kinds of material may be ejected from volcanic vents: gases, liquids consisting of molten rock, and solid material in the form of fragments; and the nature of a volcanic eruption depends largely on the proportions of these three things. If the eruption is violent and explosive, then the gases have been the chief factor in its production, and solid fragmental material is the main product; if, on the other hand, the eruption proceeds quietly, liquid rock, or lava, is the main product, and the gases play a less important rôle. We may roughly classify volcanic eruptions into those which are *explosive* and those which are *quiet.* When we classify actual volcanoes according to this difference in operation, we quickly find that, although there are good examples of both types, many, perhaps the majority, are *intermediate* in their character; that is, at times they erupt violently and at other times they quietly discharge flows of lava. In many volcanoes during a quiescent stage there appears to be a gradual accumulation of pressure, the lava rises in the conduit, and eventually the eruption begins explosively. Great quantities of gases mingled with dust and stones are ejected; and the pressure being thus largely relieved, the explosive phase is succeeded by a quieter one in which the lava escapes through rents in the cone and flows out on its exterior.

Explosive Type. — The most extreme volcanoes of this type give rise to appallingly disastrous explosions. Enormous quantities of gas are suddenly ejected into the atmosphere, so thickly mingled with comminuted rock (dust and ashes), as to form vast outrushing and expanding

Fig. 166. — Eruption of Vesuvius, April, 1906. Seen from Boscotrecase. The volcano is 4000, the ash cloud more than 17,000 feet high.

clouds of dense appearance and dark color (Fig. 166). The greatest known explosion was that of Tamboro, on Sambawa Island east of Java, in 1815. Between 28 and 50 cubic miles of material were blown into the air. The ashes produced darkness for three consecutive days for a distance of 300 miles from the volcano and dust fell over an area of 1,000,000 square miles. Krakatoa, a volcano in the Strait of Sunda near Java, exploded in August, 1883. After premonitory outrushes of gas for some time, the great explosions occurred, which blew into the air over a cubic mile of material in the form of dust and ashes. This rose as a vast dark cloud 17 miles into the atmosphere, by its denseness completely hiding the sun over an enormous area. The noise of the terrific detonations was heard as far as Australia; and the disturbance in the atmosphere was registered by barometers over the whole world.

Fig. 167. — Fiery cloud of Mont Pelée descending the mountain slope to the sea. The cloud at this moment is 7000 feet high and moving forward at the rate of over a mile in 1.5 minutes. (A. Lacroix.)

Huge waves, up to 100 feet above tide, were generated in the sea and rushed along the low-lying coasts of Java and Sumatra, sweeping far inland and destroying towns, villages, and the lives of nearly 40,000 people; these waves were perceptible 3000 to 4000 miles away.

In May, 1902, from the volcano of Mont Pelée on the island of Martinique in the West Indies, and almost simultaneously from that of Soufrière on St. Vincent 90 miles away, after small premonitory symptoms, violent explosive eruptions took place. No lava was poured out, but the intensely heated gases were so thoroughly charged with incandescent particles of rock that the heavy, fiery clouds rushed down the mountain slopes to the sea. Destroying all life in its course, the cloud on Martinique swept through the town of St. Pierre and immediately destroyed it, together with its 28,000 inhabitants. On St. Vincent 2000 people

perished and a broad tract of country was devastated. For many months after, Mont Pelée continued to eject at irregular intervals these incandescent clouds, one of which is seen in Fig. 167 rushing toward the sea.

Intermediate Type. — Probably most volcanoes belong, or have belonged, to this class. Their eruptive periods are likely to begin with explosive activity, manifested by the ejection of gases in great quantity, accompanied by solid fragmental material — bombs and ashes. In a succeeding phase liquid material issues; it may be ejected by the still issuing gases or it may break through the crater walls and produce outflows of lava, sometimes of great volume. Finally the volcano becomes quiet, its energy for the time being exhausted; the lava column sinks down in the conduit and a period of quiescence intervenes before the next eruption. Although this sketch gives in a general way the succession of events, it must not be supposed that all volcanoes of this class are alike in the character of their eruptions, or that the same one always passes through a similar set of phases at each eruption, for there is great variability in these respects. The main point is that volcanoes of the intermediate class at times are explosively active and at other times quietly emit flows of liquid lava.

Vesuvius, the longest and most studied and therefore the best known volcano in the world, belongs in this class. It occupies the site of an older volcano, which in the time of the Romans appeared to be extinct, for, although they recognized its nature, they had no traditions of its having been active. In the year A.D. 79 the volcano became active in eruptions that destroyed the towns of Herculaneum and Pompeii on its seaward flanks. A great part of the older cone on the side toward the sea was blown away or engulfed, and in its place the new center of activity, the modern Vesuvius, began to be built. This building up has continued until the present cone has become 4000 feet high. Partly enclosing it lies the sickle-shaped ridge of Monte Somma, the remains of the rim of the older crater (Fig. 168). Vesuvius is in a state of almost constant, relatively mild activity, with irregular periods of violent eruption. The last great eruption occurred in 1906 (Fig. 166).

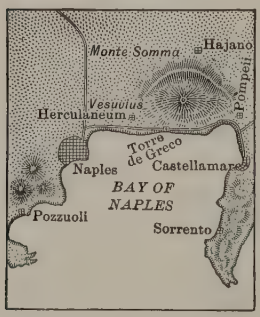

Fig. 168. — Map of Vesuvius and vicinity.

From the nature of the material composing their cones it is probable that the great volcanoes of the northwestern United States, previously mentioned, and now quiescent or extinct, belonged in the intermediate

class, as well as the active volcanoes of the Alaska Peninsula and their extension on the Aleutian Archipelago.

Quiet Type. — Volcanoes of this type give rise to quiet outflows of liquid lava, unaccompanied by the explosive disengagement of gases and the ejection of solid material as dust, ashes, and bombs. The lava of these eruptions is very hot and highly fluid. There is a more or less constant escape of gases from it, but without the catastrophic violence of the previous types. The best examples are in Hawaii.

The island of Hawaii consists of a vast mass of lavas surmounted by five cones: Kohala; Mauna Kea, now extinct (13,800 feet high); Hualalai (8300), active in 1801; Mauna Loa (13,700), now active and

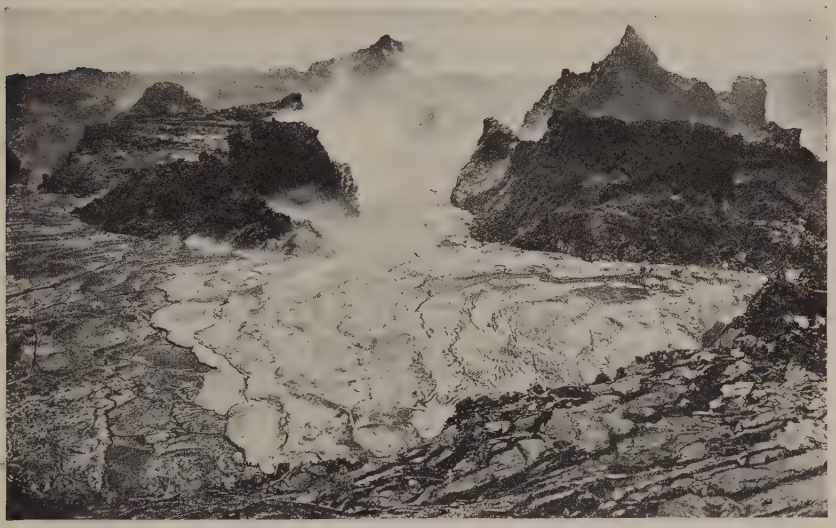

Fig. 169. — Halemaumau, north pool. October, 1921. (Hawaiian Volcano Observatory.)

some of whose lava flows have been 50 miles long; and Kilauea (4000 feet). On the eastern slope of Mauna Loa, 20 miles from its summit, is the great crater pit of Kilauea, rudely oval in shape and 9 miles in circumference. In the floor of this pit is Halemaumau, a circular depression, which before 1924 was 300 feet in diameter, and was occupied by a lake of liquid lava, boiling and fountaining from the escape of gases (Fig. 169). The temperature of the lava ranged from 1000° to 1200° C., the higher temperatures prevailing at times of increased activity. In 1924 the bottom of the pit in Halemaumau dropped suddenly 700 feet, and this subsidence was followed by explosive eruptions, the first since 1790. Avalanching from the sides of Halemaumau began and

enlarged the pit, so that it became 3000 by 3400 feet across and 1340 feet deep. After the eruption the pit filled again with lava.

Mauna Loa is the "monarch among modern volcanoes." It exceeds all others in its mighty size and in the magnitude of its eruptive activity. At times immense columns of molten lava play as fountains several hundred feet high and afford a spectacle truly sublime. Its crater is at an elevation nearly 10,000 feet higher than that of Kilauea. The outflows of lava are more likely to occur through its flanks than through the crater rim, and sometimes they take place below sea level.

Relation between Volcanoes and Magmas. — The molten rock-matter that originates within the Earth's interior and gives rise to volcanic action and volcanoes is known as *magma.* When this issues at the Earth's surface, the liquid material, and the rock produced by its cooling and solidification, is called *lava.* It must not be supposed, however, that the composition which a solidified lava shows, if determined chemically, is exactly the same as that of the magma that yielded the lava. For the deep-seated magmas contain, in addition to the mineral substances of lavas, great quantities of gases, especially water vapor, which are dissolved in them under pressure. As the magma rises to the Earth's surface and the pressure on it is consequently diminished, the gases escape, usually with more or less explosive energy, and give rise to the spectacular features of volcanic activity. As the different types of volcanoes and of the lavas that they yield depend in large measure on the magmas producing them, it is necessary at this point to consider the nature and composition of these magmas.

Composition of Magmas. — As indicated above, the substances that compose a magma may be divided into two classes: 1, those which when hot are *volatile* and which mostly escape as gases and vapors, such as steam, carbon dioxide, hydrochloric acid, sulphurous vapors, during the congealing of the lava; 2, those which are *non-volatile* and remain to form the essential ingredients of the solid lavas. These fixed constituents are *silica* (SiO_2) and the oxides of six metals, *aluminum, iron, magnesium, calcium, sodium,* and *potassium.*

Silica is an important constituent of all magmas, and metallic oxides occur in all of them, but the particular metallic oxides present in different magmas range from almost nothing to considerable quantities. However, a kind of general rule governs the composition of magmas; without going into details, which are given in the chapter on igneous rocks, it may be said that the magmas, although forming a continuous chemical series, can be divided into two classes: one in which silica and the alkali metal oxides — soda and potassa (Na_2O and K_2O) — predominate, and the other in which, conversely, lime (CaO), iron oxides

(FeO and Fe_2O_3), and magnesia (MgO) predominate. Magmas of the first class, on cooling slowly, crystallize into a mass of mineral grains that consist chiefly of alkalic feldspar,[1] with which quartz is commonly associated; the resultant rocks are called siliceous lavas. Rhyolite is the most abundant of the siliceous lavas. Magmas of the second class yield on cooling little or no alkalic feldspar, but abundant lime, iron, and magnesia minerals, such as pyroxene, calcium feldspar, and magnetite; rocks of this kind are termed basic lavas. As a rule the siliceous lavas are light-colored, whereas the basic lavas are very dark or black, and heavy because of their content of iron minerals. The siliceous lavas are sometimes termed *acidic* lavas because of the predominance of the acid-forming radicle (SiO_2) in them; the *basic* lavas are so termed because of the predominance of the bases (lime, iron, and magnesia) in them. Basalt is by far the most common of the basic lavas.

These characters and relations may be summarized in the following table:

Magmas consist of
- *a.* Volatile substances: gases, e.g., water, CO_2, etc.
- *b.* Non-volatile substances: constituents that form the solid materials of lavas.

	Chief constituents	*Resultant rock*	*Chief minerals produced*
Magma *a.*	Much silica; abundant alkalies	Siliceous lava (light-colored)	Alkalic feldspar and quartz
Magma *b.*	Less silica; abundant lime, iron, magnesia	Basic lava (dark-colored)	Pyroxene and calcium feldspar

Relation to Volcanic Eruption. — The siliceous lavas, or rather the magmas that produce them, are, when their volatile constituents have escaped, thick viscous liquids, even at very high temperatures, as high as 2000° C. as experimentally ascertained. Parenthetically, it may be remarked that 2000° is far above the temperatures that prevail at existing volcanic vents, which are generally between 1000° and 1100° and rarely reach 1200°. The extreme viscosity is due chiefly to the high percentage of silica they contain, the amount in some kinds being as much as 75 per cent of the whole. For this reason when siliceous magma rises into the upper part of the conduit where the pressure is small and the contained gases begin to escape, the magma becomes stiffly viscous and the remaining gases can escape only with difficulty and usually with violence, giving rise to explosive eruptions. Hence volcanoes that yield siliceous lavas are likely to be of the explosive type, as Mont Pelée. On the other hand the basaltic magmas, with about 50 per cent of silica, are very much more fluid, and they remain quite fluid down to

[1] For description of these and other minerals see Appendix A.

much lower temperatures, probably down to 1000° C. or lower: the gases escape from them readily, but without explosive violence, as illustrated in the lava lake of Kilauea in Hawaii. Hence the basaltic magmas usually cause quietly eruptive volcanoes.

The above statement indicates the general rule; it does not mean that basaltic volcanoes never have explosive eruptions, for a basaltic magma may become cooled in the conduit and in consequence be viscous, and thus permit its magmatic gases to escape only with difficulty and explosive energy. The explanation applies chiefly to the two extremes and indicates what is probably the most effective cause for the explosive and quiet types of volcanoes. The intermediate type of volcano may be due in part to the intermediate kind of magma erupted from it, or to this factor combined with variations in viscosity at different stages during the eruption as well as variations in the amount of gases. Sudden accession of ground water into the volcanic pipe, as undoubtedly occurred at Kilauea in 1924, will produce an explosive eruption even in a basaltic volcano (see frontispiece).

Products of Volcanoes

Gases. — It has been shown already that the products yielded by volcanoes may be divided into three general classes: gases and vapors, solid fragmental material, and liquid lava. These products will be considered in more detail, beginning with the gaseous substances. The quantity of steam discharged by active volcanoes is immense, and is indicated by the height and volume of the cloud with which many eruptions begin. This cloud consists of the dust and ashes borne aloft by the uprushing column of gases. The great quantity of steam thus discharged into the atmosphere may give rise by condensation to heavy downpours of rain in the vicinity of the volcano; and, owing perhaps to the friction of the particles and to atmospheric disturbance, the eruptions and rains are accompanied by conspicuous electrical displays and lightning. Although the composition of the gases during an actual volcanic eruption is not directly known, and it probably varies at different volcanoes and at different stages of an eruption, it is inferred with good reason from indirect evidence that it is chiefly gaseous water, or steam. As an instance of the quantity of water that some believe is discharged, Fouqué estimated that one of the subsidiary cones of Mount Etna discharged in 100 days in the form of steam the equivalent of over 460,000,000 gallons of water.

In addition to the water, the different gases and volatile products exhaled by volcanoes make a long list. Not only are they given off from

the vent itself, but the outflows of lavas continue for weeks and even months after their extrusion to emit gases as they cool and harden. Carbon dioxide, hydrochloric acid, hydrofluoric acid, and even hydrogen are given off, and to the mixture of the latter with oxygen and its sudden combustion are sometimes ascribed the explosions in the conduit. Sublimed sulphur and various compounds of sulphur, such as sulphuretted hydrogen (H_2S) and sulphur dioxide (SO_2), are emitted by some, but not all, volcanoes.

Chlorides, especially ammonium chloride, are common at many volcanoes — in fact it was the abundance of chlorides at the Italian volcanoes that suggested the idea that the eruptions are due to oceanic water leaking into the magma at depth — but no chlorides occur at Kilauea, which is on an ocean-girt island.

Fragmental Products. — These are the volcanic projectiles, the materials blown into the air by the sudden liberation of the gases. They may be derived from the crust, or plug, of hardened lava left in the upper part of the conduit after a previous eruption, from rock material torn from its walls, or from lava ejected from the upper part of the liquid column by the violent escape of the gases from the magma. Although the lava may start on its aerial flight in a liquid condition, it generally hardens in its passage and falls in solid form. The pieces of rock and the particles of magma driven upward and solidified are of all dimensions: from dust so fine that it may float in the atmosphere for several years, to large masses of several tons in weight. According to size, they are roughly classified as follows: pieces the size of an apple, or larger, are called *blocks* if ejected as solid fragments and *bombs* if ejected as particles of still-fluid magma; those the size of a nut are termed *lapilli* (meaning little stones); those the size of a pea are volcanic *ashes*, while the finest is *volcanic dust*. The ashes and lapilli are frequently spoken of as *volcanic cinders*, and the cones made of them as cinder cones. It should be clearly remembered, however, that although these terms are used to describe the appearance of the products, the " cinders " are not products of combustion. A volcanic bomb is illustrated in Fig. 170; it is of the variety called bread-crust bomb.

Fig. 170. — Volcanic bomb, Lipari Islands.

The ejected products are in part composed of compact solid rock and in part are of a spongy, cellular, or vesicular character. This vesicular

condition is due to the fact that, while the major part of the gases is passing into the air and carrying the fragments with it, a minor part is expanding in the particles of liquid, puffing them up into the cellular forms. Although the bombs, lapilli, and most of the ashes fall in the immediate vicinity of the vent and thus help to build up the cone, the dust may be carried by the prevailing winds long distances, hundreds of miles or more, and be thus spread over an immense area. Huge quantities are discharged in great eruptions, amounting to many millions of tons (Fig. 166). Such dust showers may be very destructive to vegetation and even to animal life, but the soil ultimately yielded by them is very fertile.

Liquid Material; Lavas. — In volcanoes whose periods of eruption begin explosively, the liquid lava generally issues later, after the vent has been cleared. The volcanic cone is not a structure of great strength, and is easily ruptured, or fissured, by the explosions and the pressure of the lava column, and hence the magma is not likely to flow out over the lip of the crater, but to issue through fissures in the sides of the cone. It may even happen, especially when the cone is composed of cinders, that, unable to withstand the pressure, one side gives way and allows a flood of lava to rush out through the breach thus made.

The appearance and character of a lava stream and the material produced by its solidifying depend on several things: on the chemical nature of the magma, on its viscosity, and the extent to which it retains its dissolved gases. On the chemical composition will depend the nature of the resultant rock, whether it will be a light-colored lava or a black basalt, or of intermediate character as previously explained. On the viscosity will depend the rate at which the lava will flow, the distance to which it will flow and in large measure the appearance its surface will present. When it issues, the lava is red or even white hot. It soon cools on the surface, darkens, and crusts over. If very viscous, the under part may yet be in motion, and the crust breaks up into a mass of rough, angular, jagged blocks of rock, which are borne as a tumbling, jostling mass on the surface of the slowly-moving flow. A typical Hawaiian flow of this kind advances with " a tremendous roaring, like ten thousand blast furnaces all at work at once." When eventually the flow comes to rest, the lava sheet is extremely rough and difficult to traverse. Such lava flows in Hawaii are called *aa* by the natives.

In marked contrast other lavas may harden with smooth surfaces, which exhibit curious ropy, curved, wrinkled, or twisted and billowy forms, as seen in Fig. 171. Lava of this kind the Hawaiians term *pahoehoe*, in reference to its glistening, satiny surfaces. The difference between the two varieties of lava is determined in some way yet un-

known by differences in the physical conditions during consolidation: for a flow may begin as pahoehoe and end as aa.

Very fluid lavas move with considerable rapidity, as much as 10 or 12 miles an hour, depending on the slope, Fig. 171; as they cool and become viscous the motion may be almost indefinitely slow, the stream creeping onward, possibly, for several years.

Sometimes on slopes, after the lava has crusted over, the still liquid portion beneath may run out at the lower end, leaving long galleries,

Fig. 171. — Flow of basaltic lava running down a stream bed, the water of which is turned into steam. This lava, if cooling as seen, would have the pahoehoe surface. Hawaii. (U. S. Geol. Surv.)

tunnels, or caves beneath the crust. On some volcanic cones the natural drainage passes into these tunnels, disappears from view, and issues lower on the slopes in the form of springs. Such may be in part the cause of the springs around Mt. Shasta.

Some magmas, or lavas, when ejected are too viscous to flow; they may then pile up over the vents in great domes. Such doming is chiefly, if not wholly, confined to the siliceous varieties of lava. Domes of lava have been observed in central France, Bohemia, Germany, etc., and are thought to have been formed in this way. They probably occur elsewhere. After the violent eruptions of Pelée, in 1902, the column of

siliceous lava that filled the vent and had hardened into rock was pushed up so that it rose like a vast tower above the volcano, until it attained a maximum height of 1000 feet (Fig. 172). Gradually it crumbled into a mass of blocks as a result of continuous explosions of gases.

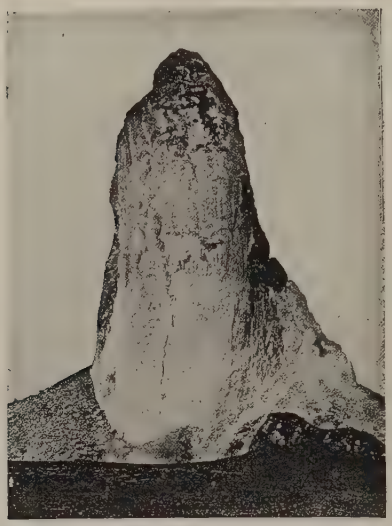

Fig. 172. — Rock tower of Mont Pelée, Martinique. 1000 feet high. (A. Lacroix.)

Effect of Contained Gases; Vesicular Lava. — That the lavas, even after they issue from the volcanic vent, still contain dissolved gases is abundantly shown not only by the clouds of steam that may issue for weeks and months from them but also by the structures they assume as they cool into stone. Thus the upper part of a flow, especially of viscous lavas of the siliceous class, may be so puffed up by the innumerable bubbles of vapor in it expanding on relief of pressure that it may become a veritable glassy froth. Such rock froth, which is usually white or light-colored, is known as *pumice*, or *pumice stone*.

In more fluid lavas, especially those of the basalt class, the bubbles are larger, and the resultant rock is spongy, cellular, or vesicular. This porous, cindery, or slag-like form of lava is called volcanic *scoria*. It is usually dark to black, or reddish (Fig. 173). Pumice, scoria, and other vesicular products are characteristic features of the upper surface of lava flows, and they constitute also a major part of the coarser fragmental materials, such as bombs and lapilli, that build up the volcanic cone.

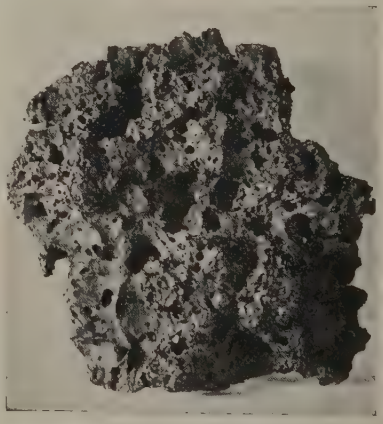

Fig. 173. — Volcanic scoria.

Consolidation of Lavas; Glass and Stone. — After lavas have been poured out and have solidified, most of them present the ordinary appearance of *stone*, but some, instead, have that of *glass*. The reason for this

is as follows. If the liquid lava is not too viscous, the chemical molecules composing it will be capable of motion and will arrange themselves into definite compounds; that is, will crystallize into mineral grains, or crystals. Possibly the crystal grains thus formed will be large enough to be seen readily and the constituent minerals will then be determinable, or they may be so minute that the lava has a homogeneous appearance; nevertheless if crystallization has taken place the lava has the aspect of stone. On the other hand, if the lava is extremely viscous or quickly becomes so through rapid cooling, the molecules may not be able to arrange themselves, or crystallize, into minerals, and the mass solidifies as a glass.

Thus while lavas in hardening into rock ordinarily take on a stony aspect, under certain conditions they may become glassy. Glasses form chiefly as the result of the freezing of siliceous lavas for, as previously explained, they are usually the more viscous. Volcanic glass is called *obsidian;* certain varieties, in allusion to their luster and appearance, are called *pitchstone.*

Some obsidians are pure glasses, others are mixtures of glass and crystals. In Yellowstone Park, Obsidian Cliff presents a section of volcanic glass 100 feet thick, which has cracked into columns in cooling. Such a thickness of purely glassy lava is unusual. It is chiefly on the edges and upper surface of lava streams that these glassy forms are found. Primitive peoples, before they gained a knowledge of metals, made much use of obsidian for making knives, arrow and spear points, etc., in a manner similar to their use of flint.

Varieties of Volcanic Cones and Craters

Kinds of Cones. — The nature of a volcanic cone depends on the material of which it is built. If composed wholly of fragmental products, the cone is high and steep in proportion to its size. This steepness is due to the high angle of repose for lapilli and volcanic ash; such material is very angular, rough, and clinging, and slopes of 40° are attained before the accumulating mass begins to slide. Cones of this kind are called *cinder cones,* and they are characteristic of volcanoes of the explosive class (Fig. 174). In contrast with them are the *lava cones,* formed entirely by quietly outflowing liquid lavas, like that of Mauna Loa (Fig. 175). They are necessarily very low and flat in proportion to their size, the angle of inclination being less than 10° (Fig. 175). These lava cones are built up by volcanoes of the quiet type. Most volcanoes, however, and this includes most of the largest volcanoes in the world, are of the intermediate type in their eruptive activity, and in conse-

quence their cones have forms that are intermediate between those just described. For they are built up by the fall of ashes and lapilli when they are explosively active and by lava flows when they erupt quietly. The great cones of the Pacific States — Mts. Shasta, Hood, Rainier, and others of the Cascade Range — have been built in this way.

The eruptions that break out on the lower flanks of the larger volcanoes give rise to smaller, or "parasitic" cones. Mt. Etna is sur-

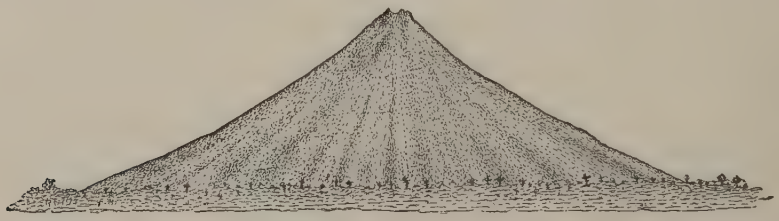

Fig. 174. — Cinder cone, showing steep angle of repose of lapilli. Outline as given by a photograph of Mayon volcano in the Philippines.

rounded by over 200 of these, some of which are nearly 700 feet high. San Francisco Mountain in Arizona, an extinct and partly eroded volcano, has a number of such minor cones, some of them remarkably well preserved. As an active volcano grows, the earlier parasitic cones may be buried and concealed under later accumulations, or, in the declining

Fig. 175. — A lava cone, to show contrast with Fig. 174. From the Snake River plain, Idaho.

stages of activity, the eruptive energy may show its last efforts in building them, as appears to have happened at the San Francisco volcano, just mentioned.

Calderas; Explosion and Subsidence Basins. — The term caldera, from the Spanish for caldron, is applied to crater-like basins of great size, especially those which are very broad as compared to their depth. The name is taken from the huge pit in the Canary Islands, called La Caldera, which is from 3 to 4 miles wide and bounded by lofty cliffs 1500 to 2500 feet high, except on one side where the encircling wall is breached. From without, as seen from a distance, the general aspect of La Caldera is that of a huge cone truncated far below its apex.

Many such great calderas occur in various parts of the world, and a

study of them has led to the view that some of them were caused by gigantic explosions that have blown away a great part of the original cones as dust and ashes, leaving the calderas to mark their sites. Perhaps more generally they have been produced by the subsidence of the column of liquid lava in the conduit of the volcano, leaving a great cavity into which the superstructure of the cone has subsided. The truncated remnant of the cone makes the rim of the caldera. Probably some calderas were formed by a combination of these two processes.

Thus Tamboro in its explosive eruption of 1815, previously alluded to, blew away a good part of the original volcano and produced a caldera nearly 4 miles in diameter.

The finest caldera in the United States is at Crater Lake in southern Oregon. This marvelously beautiful lake occupies a caldera at the

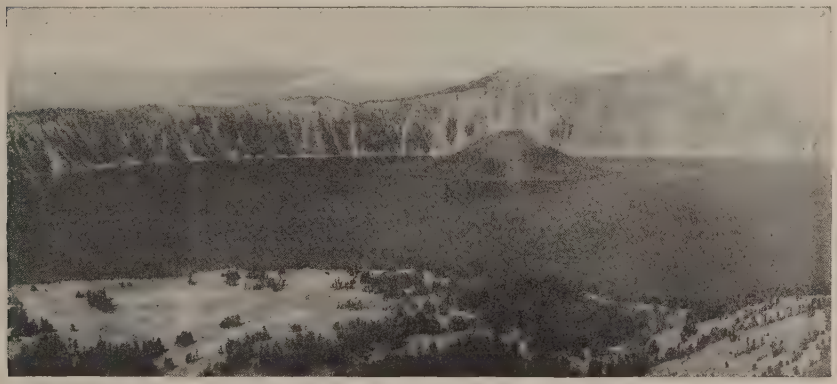

Fig. 176. — Part of the basin and wall of Crater Lake, Oregon. Note the small cone within (Wizard Island). (U. S. Geol. Surv.)

summit of a volcanic mountain in the Cascade Range, and is 6 miles long by 4 broad, 2000 feet deep, and encircled by steep cliffs 500 to 2000 feet high (Fig. 176). An island in it, made by a small but perfect cone of volcanic material, indicates a feeble renewal of eruptive activity after the principal subsidence. The caldera, if emptied of its water, would appear as a great basin. The reason for believing that the caldera was formed by the collapse and engulfment of the greater part of a former cone as the result of the subsidence of the lava column in the pipe of the volcano rather than by explosion lies in the absence of the débris — about 18 cubic miles of material — that so gigantic an explosion would have spread over the adjacent outer slopes. The former mountain, to which the name Mt. Mazama has been given, is conceived to have had the size and general character of Mt. Shasta and to have been heavily capped with snow and glaciers during the Glacial Period.

Many craters and calderas of extinct or dormant volcanoes are filled with water, giving rise to lakes. Several of the circular lakes of Italy surrounded by ejected volcanic material, like Bolsena and Bracciano, are regarded by some geologists as marking the sites of great calderas.

Explosion Pits. — In some places where volcanic activity has begun it has proceeded no further than the initial explosions that drilled a vent through the country rock. The material blown out has made a low slight ridge around the pit, but no real cone was built. Sometimes volcanic products, such as pumice and cinders, are mixed with the fragments of the country rocks. Such basins range from a few hundred feet to several miles in width, and in humid regions they are usually filled with water and form lakes. Some of the best examples of them are in the region west of the Rhine in Germany, known as the volcanic Eifel. They are called *maars* (German, *maaren*), like the Pulvermaar. A pit that strongly resembles a maar exists at Coon Butte in Arizona. The basin sunk in the plain is $\frac{3}{4}$ of a mile in diameter and 500 feet deep. The presence of meteoritic iron in and about it and other features have led to the view that it was caused by the impact of a huge meteorite and is probably not of volcanic origin. Hence it has recently been renamed Meteor Butte.

Rebuilt Volcanoes. — Not infrequently after a caldera has been formed, either by subsidence or by explosion, or both, a revival of volcanic activity starts building up a new cone within it. This is shown on a small scale at Crater Lake, where Wizard Island represents the unsubmerged top of a new volcanic cone that stands on the floor of the caldera formed by the collapse of Mt. Mazama. One of the best examples is at Vesuvius, which has built itself up within the caldera of Monte Somma, as explained previously. From this rebuilding within an old crater or caldera there results a cone-in-crater structure, of which there are many examples. The vast crater-like pits that are so common on the surface of the moon frequently show this arrangement, suggesting an analogous origin for them. Conceivably Vesuvius, before it becomes extinct, may go on increasing in size until the old caldera is obliterated; it would then be a completely *rebuilt volcano.*

Structure and Dissection of Volcanic Cones

Structure of a Composite Cone. — If a column of magma is forced upward through the crust of the Earth until it reaches the surface, the relief of pressure will enable it to commence discharging its dissolved gases and vapors. Conceivably the pressure of the contained gases may be too great for the topmost layers of the bedrock to restrain them until

the magma reaches the surface; consequently these layers may be blown into the air and a vent drilled ahead of the rising column of lava. Arrived at the surface, the magma may flow out quietly, or, if it is too viscous to do this, explosions may continue and material be blown upward. By the falling of the fragments around the vent a cone is built up, somewhat as seen in the diagram, Fig. 177. The pieces cannot, of course, fall back against the uprushing column of gases and cover the vent; they must fall outside of it, the heaviest and largest first and nearest to it, the smaller and lighter later and farther away, the distribution of the lighter depending much on the wind. Thus the cone grows as a circular ridge upon whose crest most of the ejected material is deposited. This material tends to roll and slide both outwardly away from the center and inwardly down the crater toward the vent. This process forms the cone and crater, and certain features of their structure follow as a consequence of this mode of formation.

Fig. 177. — Ideal section through a volcano. The dark layers in the cones are either buried flows or injected masses (dikes and sills).

Tuff and Breccia. — The deposits of successive eruptions are marked by layers, some of coarser, some of finer material, in each of which, if not composed of uniform-sized particles, there is a gradation from coarser at the bottom to finer at the top. Thus there arises a rude stratification, or bedding, the beds sloping down and out from the crater edge (Fig. 178). The bombs, lapilli, and ash composing the beds gradually become compacted by their weight and by the infiltration and deposition of cementing substances. They are thus transformed into a more or less friable, porous rock called, when composed of the coarser materials, *volcanic breccia*, and when of the finer dust and ashes, *volcanic tuff*. In the crater the fragments are larger, generally large blocks of rock, and they usually form a tumultuous mass without order or arrangement, with intermingled finer material; such material is called *volcanic agglomerate.*

Lava Flows and Dikes. — In addition to the beds of tuff and breccia, liquid lava flows down the outer slopes of many volcanic cones, and when these streams harden into solid rock they protect the softer layers of tuff and breccia from erosion and give strength to the edifices. Since the lava rarely flows over the lip of the crater, but, especially in high cones, breaks through fissures on the sides of the cone, these fissures also

become filled with lava, which hardens into rock. These rock-filled fissures are called *dikes*, and like ribs they also serve to strengthen the volcanic cone. Thus a vertical section through a volcano (Fig. 177) shows a central core of igneous rock surrounded by beds of tuff and breccia with intercalated flows of lava, which are cut by a radial system of dikes. This description gives us an idea of the general structure of a typical composite cone, one formed by the intermediate type of vol-

Fig. 178. — Inclined beds of volcanic ash. Part of a former cone at Trinchera, Colo. (U. S. Geol. Surv.)

cano; there are of course many variations from this, as may be inferred from what has been previously stated.

Dissection of Volcanoes. — At every stage of its existence a volcano is subject to the agencies of erosion and weathering, which tend to cut down all prominences on the Earth's surface. Its height and appearance at any given time are the result of the balance between these destructive forces and the upbuilding power of volcanism. Even active and growing volcanoes are commonly trenched and scored by ravines and gulches. After eruptions, when the cones are covered with fresh deposits of dust and ashes, the latter become so saturated with water from the rainfall that they slide down as flows of liquid mud, leaving ravines, which are enlarged by subsequent storms (Fig. 179).

As soon as a volcano becomes extinct, the ravages of erosion are unchecked, and the period of dissection ensues. The lighter tuffs and breccias are carried away more easily and rapidly; the harder, more compact, and resistant flows and dikes of lava, and the parts protected by them, are eroded more difficultly and slowly. It is surprising, however, for how long a time some cones that are composed of mere cinders

Fig. 179. — Vesuvius in 1906, showing trenching by ravines in the ashes after the great eruption. (F. A. Perret. Courtesy of Harper's Weekly.)

loosely piled will resist erosion and retain their form. The reason for this resistance to erosion is due to the porosity of the material, which allows the rainfall to sink through it without causing downwash.

As erosion progresses, the mass of rock formed by the solidification of the magma in the central conduit is brought to view, provided the magma column was not drained off before the volcano became extinct. If the central column of magma was drained off, the site of the vent probably is marked by a mass of agglomerate. As erosion progresses and the cone is demolished, the central rock mass, owing to its greater resistance, is likely to form a decided prominence, and, even when erosion has finally swept away all external evidence of the cone and bitten deeply into the underlying rocks, it may remain projecting, a monument to the vanished volcano. A rock mass of this kind filling a former conduit is termed a *volcanic neck* (Figs. 180 and 193). Here we must pause, for this carries the dissection of the volcano to its very root.

Fig. 180. — Section through a partly eroded volcano, with volcanic neck (*b*) left projecting. Broken lines show ideal restoration of cone and conduit.

Extinct volcanoes occur in every quarter of the globe and in many regions where volcanic activity has long since disappeared. Every stage of dissection is represented among them: from cones only slightly worn

to those so thoroughly eroded that the original shape has been entirely lost, but whose central rock core, outlying concentric masses of lavas, tuffs and breccias, and radial dikes still plainly show their former existence. The Rocky Mountains, once a theatre of active volcanism, in many places are strewn with the wrecks of former volcanoes. Many occur in the Yellowstone National Park and surrounding country, where, as we shall see later, the spark of volcanism still lingers. So deeply eroded are the volcanoes that their remnants now form a region of most varied and irregular topography.

Volcanoes and Deep Masses of Magma. — In the preceding paragraph there has been sketched the structure of the volcano down to a conduit filled with magma derived from below. We cannot leave the subject of volcanic activity and the structures it gives rise to before stating that volcanoes are only one phase, the surface manifestation, of the general movement of masses of magma from unknown depths into the upper region of the Earth's crust. Although these masses of magma may attain the surface and there produce volcanoes or lava flows, many of them have not reached the surface, but have remained in depth intruded into the rock layers of the crust, and there cooling and solidifying, gave rise to rock bodies of varied shapes and of sizes from a few feet in thickness up to miles in extent. Such deep bodies of magma form what are known as intrusive masses of igneous rock, and every extinct volcanic conduit, if it could be traced downward, would be found to join some such intrusive mass below, or, if active, to extend into a body of magma which in time will solidify as an intrusive mass. A proper understanding of the intrusive masses demands a more extended description and explanation, which will be found in the next chapter.

Life and Distribution of Volcanoes

Age of Volcanoes. — The span of life of active volcanoes differs greatly in different individuals. We know from written testimony that Etna has had the same general eruptive character for the last 2500 years. We estimate that because of its great volume it has taken at least 300,000 years to build this grand volcano. It is certain that the eruption of such vast masses of material as make up the larger volcanoes must have required, from the human standpoint, an immense lapse of time. On the other hand, we know from various considerations that the present active volcanoes are from the geological standpoint recent affairs.

It is also difficult to say whether a volcano is extinct, because long periods, hundreds of years, may elapse between eruptions. In the

Middle Ages, Vesuvius had been so long dormant that its crater was overgrown with vegetation and it gave no sign of life. But in 1631 it became violently eruptive, and has since been intermittently active.

New Volcanoes. — Within the period of recorded human knowledge a number of volcanoes have begun their existence, and many of them are still active. Vesuvius is, of course, the most noted case of this, but other examples are Jorullo in Mexico, which came into being Sept. 28, 1759, in the midst of a cultivated plain, and is now about 4300 feet high, and Izalco in Salvador, which began in 1770, has been almost continuously active since then, and is now over 6000 feet high.

No well authenticated volcanic eruption was witnessed within the limits of the United States proper until May, 1914, when explosive eruptions, which have since continued at intervals, began at Lassen Peak in northern California. The eruptions have been chiefly of gases, ashes, and stones. In 1915 there were two enormous explosions — two horizontal blasts that laid waste a wide swath of country. Lassen Peak, as already mentioned, is the only active volcano within the continental United States.

Distribution of Volcanoes. — The present active vents are about 430 in number, and those cones which because of their slightly eroded condition may be considered dormant or only recently extinct amount to several thousand. They have a general tendency to be grouped in long belts on the Earth's surface. The most marked of these belts is the great zone that borders the Pacific Ocean; it passes northward along the Andes, through Central America into Mexico, through the United States and Canada to Alaska, then along the Aleutian chain to Asia, and turning southward through Kamchatka, Japan, and the Philippines, it crosses the East Indies, and by various island chains again passes into the Pacific. Certain portions of this belt, like the Andes and the Aleutian chain, are remarkably linear and well developed. Another great belt has an east and west direction: from Central America it extends through the West Indies; it then continues through the Atlantic by the Azores, Cape Verde, and Canary islands, runs through the Mediterranean, through Asia Minor and Arabia, and continues along the chain of the East Indies, where it crosses the circum-Pacific belt, and extends out into the Pacific. This linear arrangement occurs not only on a large scale, affecting series of volcanic groups, but it occurs on a small scale as well, influencing the distribution of the volcanoes that compose the individual groups (Figs. 181 and 182)

Volcanoes occur both on the continents and in the oceans, the true oceanic islands appearing to be entirely volcanic. Notably in the Pacific there are great numbers of them, many extinct or dormant, some

still active, and here again many of the volcanoes are grouped in lines and stand on the submarine ridges which rise from the ocean floor. From the fact of linear arrangement has been drawn the important deduction that volcanoes are in general situated on, or near, lines of fracture, folding, and weakness in the Earth's crust.

Lines of fracture and weakness have undoubtedly proved favorable sites for volcanic action, not only for a time, but in places for long-

Fig. 181. — Map showing distribution of active or recently extinct volcanoes in the Eastern Hemisphere. On S. L. Penfield's stereographic projection.

continuing geologic periods, and thus they have greatly influenced the origin, situation, and arrangement of volcanoes. But, on the other hand, it seems clear that a volcano, or a group of them, may originate where no definite connection between them and any fracture line can be shown to exist. And in places no tendency to a linear arrangement in the group may be seen. The volcanic forces appear to have been sufficiently powerful to find an outlet without needing the aid of a fracture.

A good example of this may be seen in the Highwood Mountains, a group of extinct and greatly eroded volcanoes situated on the great plain of central Montana. While the remaining tuffs, breccias, lava flows, and dikes composing this group and their arrangement and attitudes indicate clearly the cones that once existed, erosion has dissected them so deeply that the shapes of the cones have been destroyed, the central conduits now filled with the massive rock are exposed, and

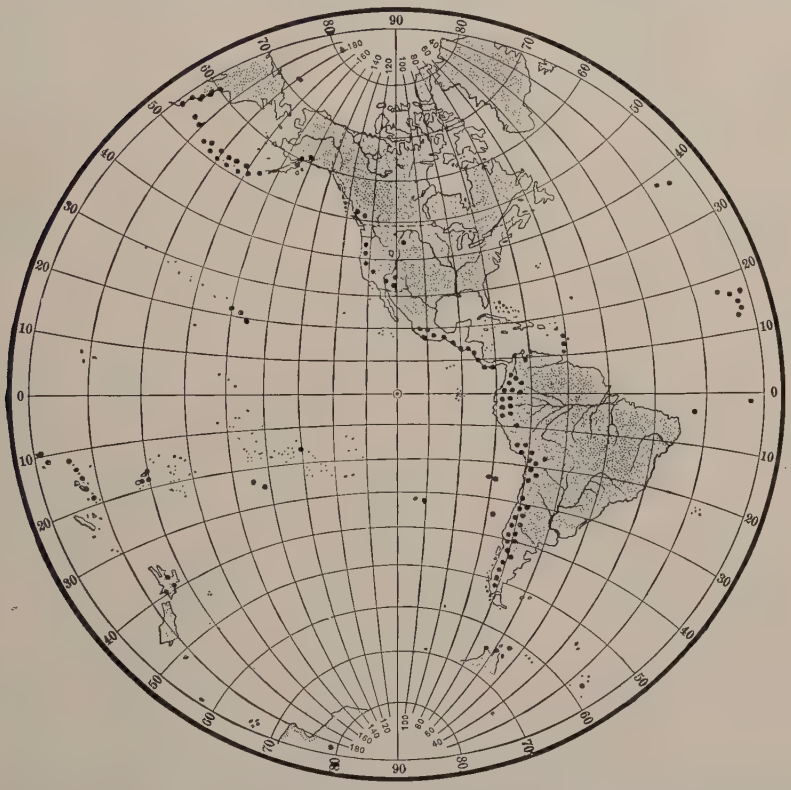

Fig. 182. — Map showing distribution of active or recently extinct volcanoes in the Western Hemisphere. On S. L. Penfield's stereographic projection.

their relations to the sedimentary bedded rocks through which they were forced laid bare. The crust shows no evidence of profound breakage or displacement that might have determined the positions of the vents, nor do the conduits of the different volcanoes show linear arrangement.

A striking instance of how little influence surface topography may have in determining the site of volcanic action, which in the immensity of its power appears to disregard such minor considerations entirely,

can be seen at the Grand Canyon of the Colorado. Uninfluenced by its 5000 feet or more of depth, volcanoes have broken out upon its very rim, instead of in its depths, and their lavas have flowed down into it.

That almost all active volcanoes are either situated in the sea, or in a general way around its borders, and when inland are in or near lakes, has led many to believe that there must be a necessary connection between the surface waters and the cause of volcanic activity. This question will be considered later in discussing the origin of volcanoes.

Submarine Eruptions. — From the great number of volcanic islands in the sea, the real oceanic islands being of this nature, it is evident that in times past tremendous eruptions and vast outpourings of lava have occurred on the sea floor. The volcanic chain of the Hawaiian Islands is an example of this. Actual eruptions beneath the sea have been observed and recognized by the issuance of vapors and ashes from the water. Thus, in 1831 a volcano was thrown up in the midst of the Mediterranean Sea, forming a new island called Graham's Island. Being composed of light cinders, it was soon destroyed by the waves and reduced to a shoal. The three Bogoslov volcanoes of the Aleutian chain formed in 1796, 1883, and 1906 are other examples.

Such eruptions have occurred repeatedly in the past and their products, mingled with sediments from the land, have been laid down as deposits on the ocean bottom, as seen in many places where the sea floor with these deposits has since been raised and become a land surface. Nor do these volcanic products differ essentially from those which are formed by volcanoes on the land. It also is probable that many of the cones formed beneath the sea, and thus protected from erosion, are of great age, even quite old from the geological standpoint, and have served as the foundations for coral islands, as previously discussed.

Fissure Eruptions. — Outflows of basaltic lava have taken place in several regions on such a gigantic scale and deluged such immense tracts of country that they cannot be referred to the outpourings of any volcanic cone or group of volcanoes. Moreover, the cones from which they might have come are apparently wanting. These great lava floods have issued from fissures in the Earth's crust. The result of such flooding is that broad plains, or plateaus, consisting of successive level sheets of basalt lava, in places interlaid with beds of tuff, have been formed.

Basalt plateaus of this origin are the great lava fields of the Columbia and Snake rivers in the far Northwest of the United States, which cover from 200,000 to 250,000 square miles and are in places 4000 feet thick; the Deccan traps (basalts) of western India, which are at least 200,000 square miles in extent and reach a maximum thickness of 6000 feet; the northern British Isles, which in part, with the outlying island groups,

appear to have been carved by the sea from a great basalt plateau that may have extended to Iceland. The horizontal layering of these lava sheets is evidently due to the extreme fluidity of the issuing magma, which permitted it to flow for many miles before congealing. Thus in Iceland, a lava flow has been traced a distance of 60 miles. It is such outpourings, which occur in other regions as well as in those mentioned, that exhibit to us the grandest effects of volcanism. It is conservative to say that since a comparatively recent geologic period as much as 500,000 cubic miles of molten material have been transferred from the inside to the outside of the globe by the extrusive process of volcanism, most of it by fissure eruptions.

Areal Eruptions. — The theoretical possibility has been pointed out that eruptions might occur through great openings formed in the crust by the foundering of the roofs of large magmatic reservoirs. Lava might thus well out on a stupendous scale. Some examples of this mode of eruption are thought to occur but have not been definitely established.

Origin of Volcanoes

General Remarks. — So far as regards the nature of volcanoes, the character of their eruptions and of the products afforded by them, their distribution, and in some measure their life, we are dealing with ascertained facts. We know also quite clearly the reason for the different kinds of eruption and the three types of cones. But when we seek to learn the cause and origin of volcanism we must then consider the depths of the Earth itself, about which we know very little. We are led from facts into almost pure speculation, and this should be clearly understood by the student. It is evident that our ideas of the cause of volcanic action will depend on those which we have concerning the nature of the Earth's interior; what has been learned regarding the Earth's interior will be considered in a later place. There are, however, certain phases of the problem of volcanism that may be considered here.

Problems of Volcanism. — Some important questions that arise when we seek to discover the cause of volcanism may be stated as follows: What is the origin of the *heat* that keeps the magmas in a molten condition? What is the origin and history of the *magmas* that come to the surface? From how deep down do these magmas come, and where is the seat of volcanism? What is the origin of the *gases;* have they always been contained in the magma, or have they been absorbed from outside sources, and, if so, when and where? And finally, what causes the magma to *ascend* to the surface from the depths and thus give rise to volcanoes? These are fundamental problems, most of which our knowl-

edge at the present time is too limited to enable us to solve. Nevertheless, the views held regarding them may be briefly stated and discussed.

Origin of the Heat. — At the present time the most prevalent view regarding the source of the heat necessary to produce the magmas is that it is original, residual from a globe once molten and still intensely hot in its interior. Some, however, regard the heat as due to the gradual contraction and compression of the Earth through the force of gravity.

Some have urged that the crushing together of the Earth's outer shell through contraction must generate heat on an enormous scale. Such compression and crushing have taken place during the formation of mountain structures, as we shall see later, and it is inferred that through this process melting has occurred and volcanoes have been made. There are two objections to this view. The first is that many volcanoes occur where there has been no crushing of the outer crust, or at least, not for an immense period of geologic time antedating the appearance of the volcanoes, as at San Francisco Mountain and other volcanoes on the high plateau in Arizona. The other is that the folding and compression of the Earth's crust that makes mountain ranges is probably a very slow process, and although great quantities of heat would undoubtedly be generated, it has not been shown why it would not be as rapidly dissipated or absorbed in doing chemical work. How could it become accumulated and concentrated sufficiently to produce melting and volcanoes? For, to use an illustration, what we need is not a cask of warm water, but a cupful of boiling water.

Inasmuch as the discovery of radioactivity has shown that some elements are spontaneously disintegrating and breaking down into other substances, as for example, uranium into helium and lead, with the concurrent production of heat in notable quantity, it has been suggested that the changes of this nature that are going on within the Earth produce the heat necessary for volcanic action and even for such vast-scale igneous manifestations as are represented by the enormous intrusive bodies beneath the surface.

Origin of the Magma. — This is a complex problem. Different volcanoes erupt lavas of different kinds. Vesuvius erupts one kind, Etna another. Furthermore, many a long-lived volcano has erupted a variety of lavas. For example, Lassen Peak, which is regarded as an old and dying volcano, has erupted rhyolite, dacite, andesite, and basalt. Did the supply pipe of Lassen Peak tap four different reservoirs containing four different magmas? It is believed that originally there was a homogeneous magma beneath Lassen Peak and that by a series of internal changes this magma yielded the other magmas. This process by which a magma of initially homogeneous composition splits

up into unlike fractions is called magmatic differentiation. It is to this process that appeal is made to account for the diversity of igneous rocks, not only at Lassen Peak but at all other volcanic centers. The causes that produce magmatic differentiation are many, but need not be explained here. It is thought that the initial primitive magma the world over is basaltic in composition. One of the cogent reasons for this belief is that the great fissure eruptions that have occurred in such enormous volumes at intervals throughout the whole span of geologic time are of basaltic composition; and this is held to indicate that a basaltic substratum of potentially liquid magma underlies everywhere the visible crust. The prevalent view is that the magma is a remnant of the original molten substance of the globe. Those who hold this view do not claim, however, that it has necessarily always been in a liquid condition. In melting, rock material expands; if sufficient pressure be put upon it, it cannot expand and, therefore, cannot melt. It is assumed that because of the tremendous pressure reigning in the Earth's depths the material although very hot is solid, but should the pressure be relieved at any place, as for instance, by upward buckling of the Earth's crust or by reduction of the superincumbent load as the result of deep erosion, or by both, then melting would ensue and a body of magma would be formed

Origin of the Gases. — The chief magmatic gas is water, as a rule making up more than 80 per cent of the total. This water may have been part of the original substance of the Earth, in short, of primitive origin — " telluric water " (from Tellus, the Earth) — or formed from the combustion of primitive hydrogen, or it may have been atmospheric water absorbed by the magma from the surrounding rocks, or it may have been acquired by the magma by melting up rocks containing water-bearing minerals. A quantitative evaluation of these possibilities is beyond the present powers of science. A magma that has dissolved much limestone would become highly charged with carbon dioxide, which is one of the important volcanic gases. Such absorption of limestone would generate enormous pressures, and such local development of pressure, it has been ingeniously suggested, may have determined the sites of certain volcanic vents.

The fact that most volcanoes are situated in, or near, the sea or lakes has been considered a strong proof that the gaseous water contained in the magma has been obtained from descending surface waters. But this argument when examined loses its force. The nearness of some volcanic chains to the sea, like those of North and South America, is only relative to the size of the continental masses. Actually they are long distances inland: in South America from 100 to 250 miles and this

includes some cones still active — like Cotopaxi — which are not near any inland water body; in North America from 30 to 100 miles or more, and, although these are mostly extinct, it can yet be shown that when active there were no bodies of water near most of them.

Cause of Ascension. — The fact that the great volcanic chains are situated on those belts along which movement and disturbance of the crust have taken place is significant. For these belts are apparently zones of weakness in the crust, and have thus afforded favorable places for the upward movement of the magma and its escape to the surface. As will be explained more clearly later, the Earth's crust is divided into great blocks, or segments, and these have in times past moved up or down with respect to one another. It is noticeable in many places that where one of these great blocks, measuring hundreds or thousands of square miles in area, has sunk, this subsidence has been attended by uprise of magma, outflows of lava, and commonly by volcanic action. Examples of such crustal subsidence and concomitant volcanic activity are the depressed tracts that form the great Rift Valley of East Africa and the valley of the Rhine. The mechanism of ascension of the magma in this way is likened to the action of a hydraulic press.

The old idea that the Earth has a hot, liquid interior and that the downward pressure of the cold and solid crust collapsing on the shrinking nucleus forces this liquid out and thus gives rise to volcanoes has been completely disproved by a number of considerations and is no longer held. The independent eruptions of adjacent volcanoes in the same group, and the fact that the lava column in Mauna Loa stands 10,000 feet higher than that in Kilauea, only 20 miles away, are disproofs of this view, and others will be mentioned later.

As to the *seat* of volcanic action, or the point from which the magma may be considered to move on its upward way, this appears to differ in volcanoes from that of fissure eruptions. Seismic evidence, from the shocks attending volcanic eruptions, indicates that the magmatic hearth or reservoir tapped by the volcano is at a relatively shallow depth — a few kilometers. Fissure eruptions, however, appear to tap the basaltic substratum, which on seismic and other evidence is believed to lie at a depth of 30 kilometers or more.

FUMAROLES AND HOT SPRINGS

Introductory. — In the foregoing description of volcanoes it has been shown what an active rôle gases, especially superheated steam, play in their eruptions. But long after a volcano has ceased to be active and has passed into a dormant or dying stage these gases continue to

issue from its crater, or from its flanks, or even from places in the surrounding country. In the same way thick beds of extruded lavas continue, often for years, to exhale steam and other vapors. And, as we shall show later, it has often happened that bodies of magma have penetrated into the outer shell of the Earth without attaining the surface or forming volcanoes, and in solidifying they, like the lavas, have given off quantities of gases, which work their way through fissures upward to the Earth's surface. It is now proposed to describe the phenomena produced at the surface by such emanations. They may appear as vapors or in liquid condition; the gaseous emanations may be considered under the general heading of *fumaroles*, the liquid emanations under *hot springs*.

Fumaroles. — This word, which is derived from a Latin verb meaning to smoke, is applied to fissures or holes in the rocks from which steam and other gases escape with more or less force. The " smoke " of the fumarole is thus mainly steam. Although steam predominates, generally forming 99 per cent of the total, other gases, such as carbon dioxide, hydrochloric acid, hydrogen sulphide, hydrogen, methane, and others also occur. Fumaroles that give off sulphurous vapors are termed *solfataras*, from the Italian word for sulphur.

In addition to the substances already mentioned, fumaroles carry certain metallic constituents such as iron, copper, and lead. These metals have been rendered volatile by the presence of chlorides and fluorides in the magma and consequently are able to leave the magma and escape into the surrounding rocks. As they approach the Earth's surface they begin to react with the other fumarolic gases and are deposited in the form of metallic minerals in the fissures through which the gases are streaming. Hematite is probably the commonest mineral formed in this way. During one of the eruptions of Vesuvius a fissure 3 feet wide was thus filled with hematite in a few days. Galena, the chief ore mineral of lead, is occasionally formed at Vesuvius by the mutual action of the lead chloride and hydrogen sulphide contained in the fumaroles. Ores as it were are actually being deposited under our eyes; in fact, it was the contemplation of these phenomena on the flanks of Vesuvius that first suggested the fruitful idea that there is an intimate relation between igneous rocks and the occurrence and origin of ore deposits throughout the world.

The temperatures of the gases issuing from fumaroles may be exceedingly high. In the remarkable fumarole field known as the Valley of Ten Thousand Smokes, near the volcano of Katmai in Alaska, temperatures as high as 645° C. have been measured. A view of several fumaroles is seen in Fig. 183.

The volcanic cone near Naples known as Solfatara last erupted in 1198; since then it has been merely discharging steam mingled with sulphur vapor, and this has given rise to the use of the term *solfataric stage* when volcanoes become quiescent or are dying. Some of the great cones of the Cascade Range, like Mt. Shasta, appear to be in a solfataric stage. In Yellowstone Park the solfataric condition still prevails and fumaroles abound. Although the steam given off in fumaroles can be mostly ascribed to magmatic origin, the amount is often increased by descending surface water that becomes vaporized, either by contact with hot rocks or by volcanic exhalations, which, as already mentioned,

Fig. 183. — General view of the Norris Geyser Basin, showing hot springs and fumaroles, and the white siliceous mass of geyserite deposited by them. Yellowstone Park.

are chiefly superheated steam. This is probably the case in Yellowstone Park.

Carbon dioxide gas is given off copiously in many places where volcanic activity still abounds, and in many where volcanism has long since died out. In some places the carbon dioxide issues directly from the ground as a gas spring, and such occurrences are known as *mofettes*. Being heavier than air, in still weather it may collect in depressions near the vent, and, as it is colorless, tasteless, and odorless, such pools of gas are deadly traps for animals by suffocating the creatures that enter them. This is illustrated by "Death Gulch" in Yellowstone Park, where animals as large as grizzly bears have become asphyxiated. But the carbon dioxide is far more likely to encounter ground water on its way upward and thus give rise to a carbonated spring, which if it passes through limestone will dissolve some of the limestone and is therefore

likely to deposit carbonate at the Earth's surface. Some carbonated springs, however, derive their carbon dioxide from the action of acid on limestone.

Hot Springs

Hot springs as well as fumaroles are likely to occur in volcanic regions. There is an intimate relation between hot springs and fumaroles: in many regions as the dry season comes on some of the hot springs become fumaroles, and when the wet season returns the fumaroles become hot springs. The evident seasonal variation leads to the theory that hot springs are chiefly fed by ground water that has become heated by magmatic steam. The water circulation is thus in principle like the hot-water heating system in a house, but instead of a furnace in the basement supplying the thermal energy, magmatic steam furnishes the heat.

It is impossible in certain regions to tell how much of the water (and steam) of hot springs and fumaroles is of surface and how much is of magmatic origin. The amount of dilution probably varies in different regions. It is estimated that the hot springs at Lassen Peak consist of 10 per cent of magmatic water and 90 per cent of water of surface origin. Hot springs in the rainless arid interior of some deserts have been regarded as mostly of magmatic origin. The proof that magmatic emanations have passed into such waters is found in the presence in them of such substances as arsenic, boric acid, and other constituents in quantities and under conditions that show that they could not have been leached out from the surrounding rocks of the country. In Yellowstone Park it is probable that most of the water is of surface origin, which becomes heated in depth by the condensation of magmatic steam in it and returns in this heated condition to the Earth's surface.

While there are various types of hot springs, according to temperature or substances in solution, the most interesting are boiling springs and geysers. Warm carbonated springs that deposit travertine have been already described in Chapter VI.

Boiling Springs. — Actively boiling springs are a feature of many volcanic regions. Many of them occur in Yellowstone Park, especially in the different geyser basins (Fig. 184). They grade from pools that are hot but rarely boil, or else simmer quietly, into springs that boil strongly and steadily, and even some that boil more or less violently and with somewhat explosive energy, interrupted by short periods of repose. The latter form transitions to the geysers mentioned beyond. So long as the supply of water is sufficient to enable the spring to have an overflow it remains limpid, and it usually has a deep blue or green

color, but if the evaporation through boiling is equal to the inflow, the water is more or less turbid from particles of disintegrated rock, and eventually becomes a mass of boiling mud. The mud may be white, or variously tinted yellow, red, purplish, or black by oxides of iron and manganese, and such hot springs are called " paint pots," " mud pots," etc. The mud as it increases in amount becomes so thick and viscid as to prevent regular ebullition, and, owing to the accumulating steam pressure, the paint pot boils spasmodically and with some violence, the mud being thrown into the air and about the vent, where it collects in

Fig. 184. — "The Devil's Punch Bowl." A hot spring boiling in the cup-like deposit of geyserite it has formed. The opening is several feet in diameter. Upper Basin, Yellowstone Park. (U. S. Geol. Surv.)

considerable masses. These are known as *mud volcanoes*, or mud geysers. They usually mark a declining stage of activity in the life of a hot spring.

Geysers. — This term, from an Icelandic word meaning to gush, is applied to certain hot springs that at intervals spout a column of hot water and steam into the air. Depending on the size of the geyser and its special peculiarities, the height to which the column of water is ejected ranges from only a few feet up to several hundreds; the eruption may last a few minutes or several hours; the quantity of water discharged may be small or it may be many thousands of gallons; the jet may play steadily and continuously straight up, or it may be fitful, be composed of minor jets, or be thrown in inclined directions. The interval between eruptions may be a definite number of minutes or hours, or it may be irregular, and several days may elapse between eruptions. Each geyser has in these ways its own peculiarities. As they are boiling springs of a special kind they are not common and are

almost wholly confined to three principal regions: Yellowstone Park, Iceland, and New Zealand.

Some geysers consist at the surface of a basin, which may be several feet to a number of yards across, and rather deep. The sides and edges of the basins usually are beautifully ornamented by the deposits of silica described beyond, and they terminate at the bottom in tubes or fissures leading to the heated depths below, as shown in the diagram, Fig. 186. The tubes and basins are, except after eruptions, filled with water at or near the boiling point. In other types the geysers by their deposits have built up mounds, or cones, of silica, from a foot or two to several

Fig. 185. — Lone Star Geyser in eruption, showing cone of geyserite. Yellowstone Park. (Haynes.)

yards high, which form upward continuations of the pipes (Fig. 185). Of the Yellowstone geysers the most celebrated perhaps is the one known as " Old Faithful," which for many years after its discovery had a very regular interval between eruptions of about 65 minutes. It is now less regular, ranging from 60 to 80 minutes. This, and the decline of activity in other geysers, or springs, does not mean any immediate diminution of thermal action in this region, but only changes going on in the underground system of pipes and fissures that supply the hot water. Altogether there are several dozen fine geysers in the park, and the number of hot springs, fumaroles, and thermal vents of various kinds amounts to several thousand.

Cause of Geyser Action. — The intermittent eruptive action of geysers depends on the relation between pressure and the boiling point of water, as was pointed out by Bunsen in connection with the great geyser in Iceland. The boiling point of water under the ordinary pressure of the atmosphere at sea level is 212° F.; increase of pressure raises it, a decrease lowers it. Thus the boiling point at the bottom of a column of water will be raised by the pressure of the superincumbent column above it; as shown in Fig. 186, it will gradually rise as we follow the tube from the surface downward. If, however, the cavity or fissure is large and open, the heated water below will rise, convection currents will be established, mixing the water, so that it will have nearly, though not quite, the same temperature in different parts of the cavity, and a regular boiling spring will result. But if the tube is long, narrow, tortuous, or constricted, convection will be prevented or restrained, and the water must boil in different levels at different temperatures corresponding to the pressures. Suppose at a point 230° in the figure the boiling point is reached; bubbles of steam are formed, the column of water above is raised a little by the expansion, the bubbles of steam rise in the cooler liquid above and collapse, the column of water settles back with jarring, thudding sounds commonly heard before eruption. The temperature of the water will gradually rise until it is just about at the boiling point for each level corresponding to its depth and pressure. Finally when a sufficient volume of steam is formed in the lower part of the geyser tube, the expansion will cause some of the water in the basin or cone at the top to overflow. This overflow lowers the pressure throughout the tube, and the water at each level, being now heated above the boiling point for the diminished pressure, will immediately flash into steam, and a mingled column of steam and hot water will be driven roaring out of the pipe into the air. After the eruption is over, the system fills again by inflow of ground water into the geyser tube, and the process is repeated.

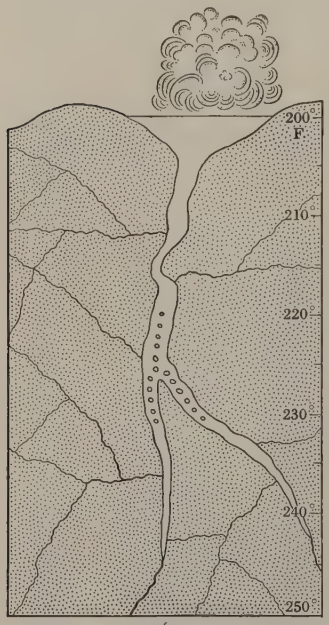

Fig. 186. — Vertical section to illustrate conditions necessary for geyser action.

The varied forms of fissures, underground conduits, and water supply account for the peculiarities shown by different geysers. It was found

by accident that adding alkaline substances, such as soap or lye, to the waters of geysers causes some of them to erupt very quickly. The government had to put a rigorous ban on the " soaping " of the geysers of the Yellowstone National Park, in order to prevent ruining them.

That the source of heat for the geysers and hot springs in the Yellowstone Park is deep-seated is shown by their occurrence in and on the shores of Yellowstone Lake, an immense body of very cold water, beneath which the rocks must be cooled to a considerable depth.

Hot-Spring Deposits. — It has been previously shown that warm springs, especially if they contain carbon dioxide in notable quantity and come up through limestone beds, form deposits of calcareous tufa, or travertine. (See Chapter VI.) But the waters of boiling springs and geysers, which occur only in regions of recent volcanic activity, are mostly alkaline and carry *silica* (SiO_2) in solution, which they deposit as a whitish material. This deposit ranges from compact to spongy in texture, and is known as *geyserite*, or more commonly as *siliceous sinter*. It forms the geyser cones, or is deposited as incrustations, much of it of great beauty, in and about the margins of the hot-spring and geyser basins. The geyser waters are dilute, in fact, so dilute as to be tasteless, and the rate of deposition is very slow when it occurs only through evaporation but is hastened by the action of organisms. Deposits of considerable size and thickness have been, and are being, made in this way, as seen forming the floor of the basin in Fig. 183. While hot springs and geysers are not geological factors that are important because of the results they achieve, nevertheless they are of great significance in a proper understanding of certain processes, such as the deposition of some ores of metals; and furthermore they are of wide popular interest.

As in the case of travertine, the deposition of silica is largely due to its secretion by low forms of vegetable life (diatoms and algae, the latter related to seaweeds), which flourish in the warm waters and even in the hot waters. The beauty of many of the pools is greatly enhanced by the rich coloring that these growths give to them.

Besides silica, the hot springs deposit other substances. The waters of some springs are acid and deposit sulphur and alum salts; and from other springs sulphides of arsenic and of metals are deposited, thereby throwing light on the processes by which ore bodies are formed.

Economic Utilization of Fumarole Fields

In recent years it has become apparent that fumarole fields may be developed so as to yield large supplies of steam for power generation. The fumarole field in the volcanic area of Tuscany, north of Rome, was

the first to be developed. Wells have been put down to depths of 600 feet, and the flow of steam, as well as the temperature of the steam, has been found to increase as depth is gained. About 30,000 horsepower is generated and is transmitted to Florence, 60 miles away, Pisa, and other cities.

At " The Geysers," in the Coast Ranges of California, 40 miles north of San Francisco, is an area of 35 acres containing a few feeble fumaroles and some small but very hot springs. The name " Geysers " is a misnomer, however, as none of the hot springs is periodically eruptive. In 1921 the idea was conceived of drilling wells in this area to develop a flow of steam for power purposes. So far eight wells have been put down, the deepest being 650 feet, and copious supplies of superheated steam have been developed. In fact, it is estimated that four of the wells will on the average deliver more than 1300 horsepower each. As in Tuscany the deeper a well is drilled the greater the steam flow and the hotter the steam.

In Java also the power possibilities of its fumarole fields are being investigated. One of the fields was bored in 1926. The most promising well, 220 feet deep, yields steam sufficient to generate 1200 horsepower. Many fumarole fields remain to be bored in Java, as well as others in the nearby islands of Sumatra and the Celebes.

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CHAPTER XI

THE IGNEOUS ROCKS

The igneous rocks form one of the great divisions of the rocks that make up the crust of our planet. As their name implies, heat was an essential factor in their origin, and they may be defined as those rocks *which have been formed by the solidification of molten matter that originated within the Earth.* Such molten matter, as was explained in the discussion of volcanic action, is commonly called *magma*, a term we shall frequently use.

Distinguishing Characters. — The features that distinguish the igneous from the sedimentary and metamorphic rocks consist partly in the relation that the igneous masses exhibit towards other rocks with which they occur or with which they are in contact (a relation that we term their mode of occurrence), and partly in the characters that become evident when the rock itself, instead of the mass of which it is a part, is closely examined.

Igneous rocks do not of course contain fossils, nor do they show as a rule the parallel or banded appearance of the stratified rocks. They have also certain distinctive peculiarities in the arrangement of their component mineral grains (the *texture*, as it is called). Some igneous rocks, indeed, are more or less made of glass, which at once betrays their origin, because glass is formed only by the chilling of molten material. The textures will be described when the different kinds of igneous rocks and their classification are considered; first we will discuss the ways in which masses of igneous rocks occur as elements in the architecture of the Earth's crust.

OCCURRENCE OF IGNEOUS ROCKS

Intrusive and Extrusive Rocks. — There are two chief modes of occurrence of igneous rocks: the *intrusive* and the *extrusive*. In the intrusive mode of occurrence the magma, at the time it was rising from the depths, stopped before reaching the surface, and consequently it cooled and solidified under the cover of the rock masses of the Earth's outer shell. In the extrusive mode of occurrence, the magma attained the surface: it was extruded upon it, and has solidified there. The extrusive rocks are sometimes called *effusive* and sometimes *volcanic*

rocks, although the term volcanic is perhaps not strictly applicable to those lavas that were extruded from fissures rather than from volcanic vents.

Although the division of igneous rocks into intrusive and extrusive is a natural one, the two classes are closely connected and, in fact, grade from one into the other. The magma of any extrusive mass came up through some passageway from below; this passageway remained filled, and eventually the magma in it solidified into rock. Consequently in theory every extrusive body that occurs on the Earth's surface is connected with an intrusive mass occurring below. In some places this connection of the extrusive mass with its root may be seen, but more generally the extrusive covers the root or has been separated from it by erosion, and the former continuity has been destroyed. It is clear also that we must think of the intrusive prolongation as extending downward and connecting with some greater mass of magma (or rock) below, of a nature to be presently described. See in this connection what has been said regarding the relation between volcanoes and deep-seated masses of magma in the preceding chapter.

Both intrusive and extrusive rocks have various modes of occurrence. The mode of occurrence of intrusive masses depends on how the intrusive mass is related to the rocks that enclose it; and the mode of occurrence of extrusive masses depends on the conditions under which the magma was ejected. We shall begin with the modes of occurrence of intrusive masses; but it should first be recalled that, inasmuch as these intrusive masses were covered at the time of their formation by the rocks into which they were intruded, they can be exposed at the Earth's surface and thus laid open to observation only after erosion has carried away the cover and has disclosed their intrusive character. In some places, where the magmas were intruded under a thin cover, the time necessary to do this may have been short; but in other places, where the igneous masses were deeply buried, it may have been exceedingly long.

Intrusive Modes of Occurrence

The principal kinds of intrusive igneous bodies are *dikes*, *sills*, *laccoliths*, *necks*, *stocks*, and *batholiths*. Several other modes of occurrence are recognized, but as they have not the importance of those mentioned, they will be treated for simplicity's sake as modifications of them. The simplest form of intrusive body is the dike and this will be considered first.

Dikes. — A dike results from the solidification of magma that has filled a fissure in preëxistent rocks. Consequently, it is a tabular mass:

its length and breadth are great compared with its thickness. It may "cut," that is, pass through, rocks of any kind — igneous, sedimentary, or metamorphic. In the sedimentary rocks it must by definition cut the planes of stratification at an angle; if, however, the igneous mass lies parallel to the bedding planes it is termed a sill. A dike may be a few yards or many miles long; it may be a fraction of an inch, or many hundreds, or even some thousands of feet thick. An illustration of a dike is seen in Fig. 187.

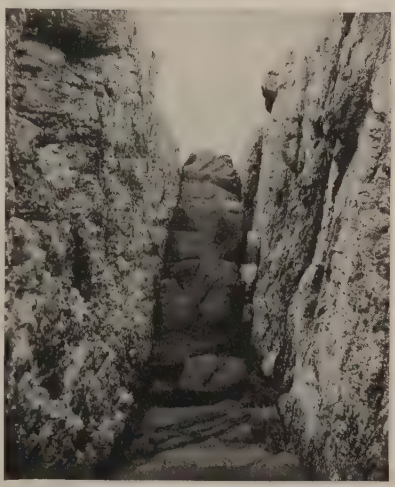

Fig. 187. — Dike of trap rock in granite. This dike is less resistant than the enclosing granite and has been cut away by erosion, leaving a trench in the granite. Isles of Shoals, N. H.

Most dikes are from 2 or 3 feet up to 20 thick; the length varies greatly. A great dike in the north of England extends for over 100 miles. The angle of inclination of the plane of extension of the dike with the horizontal is called its *dip*. The direction of its outcrop, or intersection with the horizontal plane, is termed its *strike*, or trend.

Some dikes have attained the Earth's surface and given rise to outflowings of lava, but others have not reached it and have become exposed by subsequent erosion. Some dikes were the canals that fed larger intrusive bodies above them, such as the sheets and laccoliths to be next described. In the process of erosion, a dike may be more resistant than the surrounding rocks and hence is left projecting as a wall; some, however, are less resistant and form ditches; from these features the name is derived, especially from its resemblance to the more prominent wall, for dike means both wall and ditch. The rock of some dikes is divided into blocks by joints, and very commonly the blocks are columns lying perpendicular to the walls of the dike, like a pile of cordwood, an arrangement whose origin is described later under columnar structure. Dikes occur at many places in more or less well-defined systems, and around volcanic centers are likely to be radially disposed.

Sills. — It is not uncommon to find, where magma has been intruded in bedded rocks, that it was injected as layers between the beds. Injection of this kind most frequently happens where the beds are easily penetrated, as in shales, thinly bedded sandstones, and the like. An igneous mass that thus lies concordantly between the bedding planes is

known as a *sill*, or an intrusive sheet. Sills may be a foot or so in thickness or several hundred feet, and they may be many square miles in area. An illustration of them is seen in Fig. 188.

Sills may break dike-like across the strata and then continue along a new horizon. Sills may be distinguished from flows of lava that have been buried by deposits of later sediments by the fact that the rock composing the sills is of the same compact nature at top and bottom, i.e., is not slaggy or scoriaceous like a lava flow, and by the overlying sedi-

Fig. 188. — Sills of igneous rock. Cottonwood Canyon, New Mexico. (U. S. Geol. Surv.)

ments having been baked and altered by the intrusion. The surface of a lava flow is usually spongy, ropy, slaggy, etc., and a flow could of course exert no action on beds not yet deposited upon it. Sills are most likely to occur where larger intrusions of magmas, such as laccoliths and stocks, have taken place, as accompanying features in the surrounding strata. In regions where thick sills occur and the strata have been dislocated and upturned, they may give rise to prominent topographic and scenic land features through the effects of later erosion. This is illustrated in some of the trap ridges of southern New England, northern New Jersey, and in other places.

Laccoliths. — The laccolith in its typical development is a lenticular or dome-shaped mass of igneous rock that was intruded into sedimentary rocks between the bedding planes. It has a flat floor, and is more or less circular in ground plan. If during the formation of a sill, magma is supplied more rapidly from below than can easily spread laterally

away from the supply channel, the overlying strata will be arched up, as if by a hydraulic press, and a thick lens of liquid rock will be produced, giving rise on solidification to a laccolith. Such a mass may be a few hundreds of feet or more than a mile thick at the center, and a few

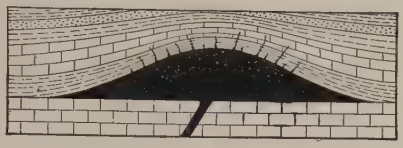

Fig. 189. — Section of a laccolith. The black area is the igneous rock.

hundreds of yards or many miles in diameter. A section of a laccolith is shown in Fig. 189, and a photograph of one from which the cover has been removed by erosion, exposing the igneous rock, is seen in Fig. 190.

While the above statement gives the idea of a typical laccolith, many departures from this arrangement are found in the actual occurrences.

Fig. 190. — Bear Butte, a laccolith denuded of its cover and the igneous mass laid bare. The ring of upturned eroded strata is seen about its base. Black Hills, South Dakota. (U. S. Geol. Surv.)

In ground plan they may be circular, oval, or quite irregular, and instead of being symmetrical in section, as in Fig. 189, they may be wedge-shaped. According to their degree of flatness, all transitions into sills occur. They may also break across the strata in places like sills. They may thin out into sills, or be accompanied by sills on the flanks of the arches, and thus be compound in structure. Such sills may themselves swell out into inclined lenticular masses, or subordinate laccoliths.

And in regions where strata were being folded, areas of relief from pressure or openings might form on the sides of the arches that would permit the entrance of magma (Fig. 191). This would give rise to inclined, doubly convex bodies like that shown in Fig. 192. Laccolithic bodies of this character have been termed *phacoliths* (from the Greek words for lentil + stone).

It is sometimes asked whether the magma itself supplies the force necessary to make room for itself by arching up the strata; in other

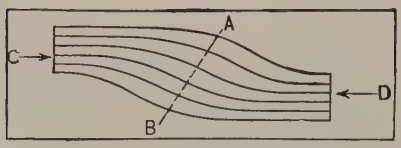

Fig. 191. — Strata being folded by compression *CD*, relief from pressure from overlying beds and spreading might occur in direction *AB*.

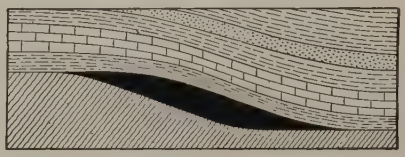

Fig. 192. — Section of an inclined laccolith, or phacolith.

words, was the magma *aggressive*, or as indicated above, was the force lifting the beds produced in some other way, and did the magma simply flow into the space that was opening for it, the intrusion having been, so to speak, *permissive?* A study of known laccoliths shows that in all probability both of these modes of intrusion occur. In central Montana, where the strata are horizontal and undisturbed save where intrusive masses occur, the magma must have acted aggressively, but in other places where folding and uplifts occur, the intrusions were probably permissive.

Laccoliths, more or less exposed by erosion, are conspicuous features in many parts of western North America, where they were first discovered by Gilbert in the Henry Mountains of southern Utah. Subsequently they have been found in various parts of the world and are therefore a not uncommon form of intrusion. Moreover, some of the more recently recognized laccoliths are of immensely greater size than the classic laccoliths of the Henry Mountains, the largest of which has a volume of 10 cubic miles. The great Duluth laccolith on the northwest shore of Lake Superior is estimated to represent the injection of 50,000 cubic miles of magma.

The floor of a large laccolith as a rule is not horizontal but dips from the perimeter inward toward a point under its center, just as if the floor had sagged from loss of support. Such loss of support is conceivably the result of the emptying of a magma reservoir below the laccolith.

Necks. — When a volcano becomes extinct, the column of magma that filled the conduit leading to unknown depths below will solidify and form

a cylindrical mass of igneous rock. Erosion will in time cut away a great part of the ashes and lavas of the cone, leaving this more compact and resistant rock projecting, as shown by the line *abc* in Fig. 180. The level of erosion may eventually descend into the rocks that form the basement on which the volcano stands; and all of the ashes and lavas having been thus swept away, only the projecting mass remains to mark the former site of the volcano. Such a mass of rock is known as a *volcanic neck* (Fig. 193). It is commonly more or less circular in ground

Fig. 193. — Alesna volcanic neck, Mt. Taylor region, New Mexico. (U. S. Geol. Surv.)

plan and may be from a few hundred yards up to a mile or more in diameter. The rocks about volcanic necks are likely to be cut by a radiating system of dikes, and commonly, if stratified, injected with sills. The significance of volcanic necks has been previously explained in the discussion of volcanoes.

Stocks. — This term is applied to certain large domal bodies of intrusive rock that in the form of magma have ascended into the upper levels of the Earth's crust and there solidified. They have become visible because erosion has stripped off the covering rocks. They have as a rule a more or less circular or oval ground plan. Their outer surface, or contact surface, cuts across the inclosing rocks, is more or less irregular, and the mass may widen in extent as it descends. Their size ranges from a few hundred yards to several miles in diameter. As they are likely to form protuberant topographic features after being exposed by erosion, they are sometimes called *bosses*. The distinction from a

volcanic neck is not one of size alone, though necks tend to be smaller than stocks, but in that the term " neck " is employed only when there is evidence that the igneous body has functioned as the supply conduit of a volcano. Some stocks were doubtless necks, but this cannot now be proved.

Batholiths. — A batholith is a huge intrusive mass of igneous rock. According to some definitions, a batholith is " bottomless," that is, it extends indefinitely downward into the Earth's crust, in contrast to a laccolith, which rests on a floor. The " bottomlessness " is of course pure hypothesis. A batholith differs from a stock only in its much greater size, as some are exposed over many thousands of square miles of surface. Arbitrarily, an intrusive igneous body less than 40 square miles (100 square kilometers) in areal extent is called a stock; if larger than 40 square miles, a batholith Some stocks, as shown in Fig. 194,

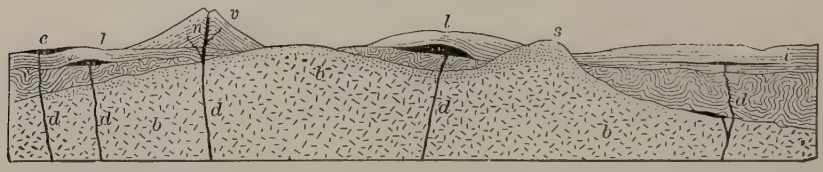

Fig. 194. — Diagram to illustrate the occurrence of igneous rocks: *b*, batholith, partly uncovered by erosion; *s*, stock, partly uncovered by erosion; *n*, volcanic neck forming *v*, a volcano with tuffs and breccias; *l*, *l*, laccoliths; *i*, intrusive sheet, or sill; *e*, extrusive sheet; *d*, *d*, dikes. The batholith and stock belong to an older generation than the other igneous bodies shown. If the section went deep enough, it might show a younger batholith from which the smaller bodies and the volcanic materials were derived. Horizontal distance shown thirty miles; vertical distance, three miles.

are merely dome-like protuberances from the body of an underlying batholith; they have been aptly called cupolas.

The largest batholith in the United States is the Idaho batholith in central Idaho, extending over an area of 16,000 square miles. The causes for the rise of so stupendous a mass of molten rock matter into the higher levels of the Earth's crust and the processes by which it makes room for itself are among the most fundamental and fascinating problems of Geology.

Extrusive Igneous Rocks

The modes of occurrences of the extrusive igneous rocks have already been described in connection with volcanoes and extrusions of lava, and what has there been said in regard to them may be profitably consulted in this connection. For the sake of convenience the following summary is here given. Magma is erupted in two ways, depending on the quantity and activity of the gases contained in it: the *quiet*, in which it wells

out as a liquid and solidifies into rock, and the *explosive*, in which it is violently driven into the air and falls in the form of solid fragments.

Quiet Eruption: Lava Flows. — Magma that reaches the surface and pours out is known as *lava*. When solidified it is commonly spoken of as a lava flow, or an extrusive sheet. Usually such flows are poured out from volcanoes. The extrusions of a few volcanoes like some of those in Hawaii, are indeed almost wholly of this nature, but generally lava flows succeed or alternate with ejections of fragmental material.

Some lava flows have not been erupted from volcanoes, but have welled out quietly from fissures. In the geologic past fissure eruptions have occurred on a huge scale, as in the Columbia River region of the northwestern United States, in western India, and in the north of the British Isles. In each of the first two regions the pile of superposed lava flows is thousands of feet thick and covers an area of approximately 200,000 square miles.

Not infrequently sheets of lava have sunk below sea level and been covered by deposits or they have been erupted on the sea floor and have been covered by sedimentation. Still more commonly lava flows on the land have been buried under various kinds of continental deposits. Such buried extrusive sheets are distinguished from sills by the fact that they have not altered or baked the sediments above them, and their upper surfaces usually show the structure common to the surface of lavas, such as the vesicular, scoriaceous, and ropy ones described previously. Furthermore the layer of sediments directly above a buried lava flow is likely to contain pebbles or boulders derived from the upper part of the flow before it was entirely buried.

Explosive Eruption: Tuffs and Breccias. — When magma attains the Earth's surface in the conduit of a volcano, it may erupt as quiet flows of lavas as already mentioned, or, if its viscosity is sufficient and it is charged with vapors under great tension, it will give rise to explosive activity, and the material will be hurled into the air. Owing to the expansion of the contained gases, chiefly steam, the ejected pieces usually are more or less vesicular, and range in size from large masses to fine dust. This material is roughly classified according to size, as previously explained.

During an explosion in a volcanic vent not only are fragments of hot, still-fluid magma ejected, but also great quantities of cold solidified lava disrupted from the crater walls are blown into the air. The coarse angular pieces produced by the fragmentation of the cold lavas are termed *blocks*, in contra-distinction to the bombs, whose roundish forms show that they were still viscous during their aerial flight. The coarser material — the blocks, bombs, ashes, and lapilli — falls around the vent

and builds up the cone; the lighter ashes and dust, carried by air currents, tend to fall after these, and at greater distances. The coarser material thus produced is termed *volcanic breccia*, while the finer material is known as *tuff*.

Tuffs and breccias are widely distributed, occurring wherever volcanic activity is being or has been displayed, and their presence is, indeed, one of the surest indications of former volcanism in places where it has long since died out. We are thus able to recognize that volcanoes formerly existed in various parts of the eastern United States and Canada, from Nova Scotia to Georgia. Tuffs and breccias occur in vast quantities piled up in places thousands of feet in thickness in the Rocky Mountains, where, as in western Wyoming, serried mountain peaks have been sculptured from them by erosion.

Age of Igneous Rocks

The geologic period when a given mass of igneous rock was erupted or intruded is determined by ascertaining its relations to the rocks with which it has come in contact. Thus, if a body of igneous rock, such as a dike, cuts through another body of rock, it is manifestly the younger of the two. If it cuts across stratified beds it is younger than they are, and lavas of course are more recent than the rocks upon which they lie. If a sheet of igneous rock lying concordantly between strata has affected the beds both above and below it (see contact metamorphism, page 352), it is younger than both. If it has not baked or otherwise metamorphosed the overlying beds, the sheet may be a lava flow and older than they are. It is thus usually easy to tell when an igneous mass is younger than other rocks by examining its contacts with them. The age of the stratified rocks is of course determined by the fossils that occur in them, and the endeavor is made to find whether the igneous rocks, which contain no fossils, are older or younger than the fossiliferous rocks with which they may be in contact.

KINDS AND CLASSIFICATION OF IGNEOUS ROCKS

Introductory. — The features by which the igneous rocks are distinguished have been already mentioned in a preliminary way in discussing the products of volcanoes, but they should now receive the attention that they demand. Igneous rocks are divided into different kinds on the basis of two properties: first, their texture; and second, their composition. Each of these properties requires explanation.

Texture. — The most obvious thing about an igneous rock is its texture. By *texture* is meant the *relative size or sizes* of the component

grains and the shape and arrangement of the grains. Thus if the grains are as large as peas, we say that such a rock is *coarse-grained* in texture; if the grains are the size of those in granulated sugar, we say that the rock is *fine-grained;* whereas, if the particles are so minute that they cannot be discriminated by the unaided eye and the rock looks as if it were a homogeneous substance, we say that it is *aphanitic*[1] in texture.

The texture depends on the *rate at which the magma cooled.* For, if the magma is too hot, as previously explained, crystallization cannot take place, and no crystals will begin to form until the temperature has fallen far enough; then they will begin to separate from the magma and, if the cooling is very slow, they will have time to grow to large size, thus producing a coarse-grained rock. But, if the cooling is rapid, more and more new centers of crystallization will be forced to form, and, if the process is thus hurried, instead of few crystals growing to large sizes, the rock will consist of a large number of smaller particles and will, therefore, be fine-grained in texture. And with still more rapid cooling the particles may be so minute that they are not discriminable by the unaided eye and the resultant rock is of aphanitic texture. Analogy will now carry us one step more: we can conceive that the cooling may take place with such great rapidity that the magma will solidify into a homogeneous substance before any crystallization, which consists in the molecules arranging themselves together to form definite solid compounds, can occur. In this event the result will be a *glass,* or a glassy rather than a stony texture, a case that is by no means uncommon.

To sum up, then, we see that igneous rocks may be *coarse-grained, fine-grained, aphanitic,* or *glassy* in texture, and that which of these textures is developed depends on the rate of cooling of the magma.

Porphyry: Porphyritic Texture. — In what has been said so far regarding the texture of igneous rocks, it has been tacitly assumed that the component mineral grains in any given rock are of uniform size, or that the rock is evenly *granular,* as it is called. Not all igneous rocks, however, are evenly granular. Inspection shows that many of them are composed of crystals of two sizes: some crystals that are larger and more distinct and which are embedded in a matrix of much finer grains. An igneous rock having this texture is called a *porphyry.* Examples of the even-granular and porphyritic textures are seen in Fig. 195, *A* and *B*. The matrix of a porphyry is termed the *groundmass,* and the large crystals

[1] *Dense* is often incorrectly used in America as a synonym for aphanitic. It is also used correctly to mean "of high density" and this double usage leads to ambiguity. A "dense" felsite (in the sense of an ultra-fine grained rock) is not a dense rock.

embedded in the groundmass are called the *phenocrysts* (clearly discernible crystals). Porphyritic rocks are common.

The groundmass may itself vary widely in grain size in different porphyries: it may be medium-grained, fine-grained, aphanitic, or glassy; most commonly it is aphanitic, as in the extrusive rocks. The phenocrysts also may vary widely: they may be of large size, as large as walnuts or as small as grains of sand; they may be abundant or comparatively few. But in all porphyries there is this *contrast* between *sizes* of crystals, between groundmass and phenocrysts, which makes the essence of a porphyry. The student should guard against thinking that the porphyritic texture is a contrast of *colors;* thus a rock consisting of

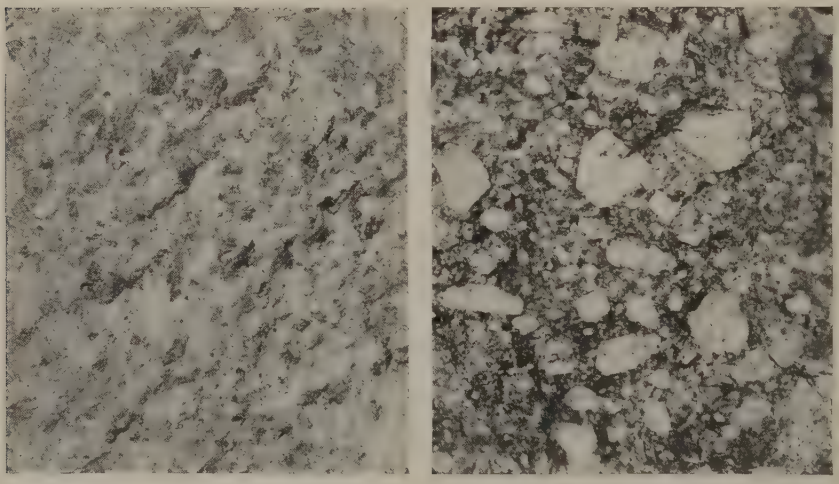

Fig. 195. — *A.* Even-granular Rock. *B.* Porphyry.

grains of light-colored quartz and feldspar, in which are embedded a few black crystals of mica, all grains being of about the same size, is *not* a porphyry.

Relation of Texture to Geologic Mode of Occurrence. — Since, as has been shown, the texture of an igneous rock depends chiefly on the rate at which the magma cools, it is clear that this rate will depend in turn most largely on the volume of the magma. Obviously an *intrusive* mass of magma that is surrounded and blanketed above by other, older rock masses must lose heat much more slowly than an *extrusive* one, which is poured out on the surface in the form of lava. Hence, as a coarse-grained texture is the result of slow cooling, we naturally associate it with intrusive masses, and, conversely, we regard the aphanitic or glassy textures as belonging to the products of extrusion — the

lava flows. But it is also clear that the size of the intruded mass will greatly influence the rate of cooling, since a very large mass cools more slowly than a small one. Thus, the rocks that make up great batholiths are coarse in texture, whereas the rocks of dikes and sills tend to be much finer-grained. On the other hand, the central portion of an extremely thick lava flow may cool with sufficient slowness to develop a medium-grained texture, whereas a magma that was forced into a narrow fissure in cold rocks might be chilled so quickly as to assume an aphanitic, or even glassy texture. Thus various modifications of the general rule according to particular cases can be easily imagined; nevertheless this general rule, that the intrusive rocks are medium- to coarse-grained and the extrusive rocks are fine-grained to aphanitic, holds true.

An important deduction that follows from the above is that the coarse-grained rocks, because they have been formed deep within the crust, can become visible at the Earth's surface only after a period of denudation that has been sufficiently long to remove the covering rocks and expose the igneous mass.

While the rate of cooling is the most important factor that determines the texture of igneous rocks, as discussed above, it is not the only one. The subject is too complex for detailed treatment in this work, but it may be mentioned that the chemical composition also has its influence. Under similar conditions of cooling, basic magmas (those low in silica and high in iron and magnesia) tend to assume a coarser grain than those composed of much silica, alumina, and alkalies. The reason is that the basic magmas are more fluid than the siliceous magmas, as already explained, and thus when they crystallize this mobility of the molecules permits the crystal grains to grow to larger sizes.

Also the presence of the included gases that magmas contain, especially the water, increases the fluidity and thereby promotes a coarser crystallization. This is very notably shown in and around certain intrusive masses by dikes that are made up of large and even huge crystals of quartz, feldspar, and mica. Crystals several feet in diameter are not uncommon. The very coarse masses of this composition are known as *granite pegmatites*, and from them are obtained the plates of mica that are used commercially. In allusion to their giant grain the pegmatites are sometimes called giant granites.

In a volcanic neck the rock is likely to be comparatively coarse-grained in spite of its small mass, because the constant upward passage of molten material to the surface causes the rocks surrounding the conduit to become greatly heated, thus producing slow cooling of the last charge of magma that occupied the conduit and solidified there when the volcano became extinct.

Composition of Igneous Rocks. — Since igneous rocks are produced by the solidification of magmas, their composition will obviously depend on the chemical composition of these molten fluids. It has been already shown (Chapter X) that a magma consists of two parts: a volatile part, consisting of the fugitive constituents — water vapor, carbon dioxide, sulphur fumes, amounting to a few per cent, and a non-volatile part, consisting of the fixed constituents, chiefly molten silicates. Although for reasons we cannot now consider, the gaseous part is important in rock formation, it is essentially the molten silicates that give rise to the rocks and are the chief constituents of magmas. While it is evident that a magma cannot be analyzed directly, still the fixed constituents and their relative proportions can be ascertained by analysis of the cold and solid rock. Many thousands of igneous rocks from all parts of the world have been analyzed and the following results have been obtained, which show the ranges in the amounts of the various constituents. These results are reported by the chemist in terms of oxides, and the convention has grown up in speaking of the chemical composition of magmas and rocks as if they were actually composed of these oxides. When we say that a rock is high in magnesia we really mean that its chemical analysis shows that a large amount of magnesia is present: this mode of expression implies nothing as to the chemical combination of the elements in the rock or magma.

Silica, SiO_2, always present; ranges from 35 to 80 per cent.
Alumina, Al_2O_3, ranges from 0 to 25 per cent.
Oxides of iron, FeO and Fe_2O_3, usually both; 0 to 20 per cent.
Magnesia, MgO, 0 to 45 per cent.
Lime, CaO, 0 to 20 per cent.
Soda, Na_2O, 0 to 16 per cent.
Potash, K_2O, 0 to 12 per cent.

It must not be concluded from the above table that any and all sorts of mixtures of these oxides can occur within the limits shown. As we shall see presently, certain general laws govern their associations. It will also be noticed that there is one acid-forming oxide (silica) present, while the oxides of the six metals (aluminum, iron, magnesium, calcium, sodium, and potassium) are in general bases. Oxides of other elements occur in small or minute quantities, but are of so much less importance that they may be neglected.

Associations of Oxides in Rocks. — Although there are many exceptions to this rule, it is generally true that large percentages of potash and soda (alkalies) in a rock are accompanied by a correspondingly large amount of silica and consequently by small amounts of the other three metallic oxides. Conversely, large percentages of magnesia, lime, and iron ox-

ides are likely to be associated, and these go with low silica, the alkalies being small or wanting. These reciprocal relations are of fundamental importance in igneous rocks, and it will be recalled that they have been pointed out before, because the nature of volcanic activity and the kinds of volcanoes in large measure depend on them. They may be expressed in a general way as follows:

Where SiO_2, $(Na,K)_2O$ are high, CaO, MgO, FeO are low or wanting.
Where CaO, MgO, FeO are high, SiO_2 is low, and $(Na,K)_2O$ are low or wanting.

Crystallization. — It is a familiar experiment that if a liquid containing a salt, zinc sulphate for example, is boiled down and concentrated to a certain point, all of the zinc sulphate can no longer remain in solution, but will begin to appear as a solid in the form of crystals. If the hot solution is allowed to cool, more crystals of the salt will be formed, since hot solutions as a rule can contain more salt than cold ones. In analogy with this, a magma can be regarded as a solution; it contains dissolved in it various salts (mineral molecules) more or less electrolytically dissociated. If the magma cools with sufficient slowness, the dissolved matter in it will separate from it as crystals. This crystallization will proceed as the temperature falls until the whole magma has turned into a mass of solid crystal grains. The molten liquid has become *stone.* The minerals separate from any given magma in a definite order, which is governed by their solubility in that magma.

Kinds of Minerals. — The more important minerals that form the igneous rocks are the following:

Feldspar Group	*Ferromagnesian Group*
Orthoclase Feldspar, $KAlSi_3O_8$	Mica (Biotite), $(H,K)_2(Mg,Fe)_2Al_2Si_3O_{12}$
Plagioclase Feldspars	Pyroxene, $Ca(Mg,Fe)Si_2O_6$
	Hornblende, $Ca(Mg,Fe)_3Si_4O_{12}$
	Olivine, $(Mg,Fe)_2SiO_4$
Quartz, SiO_2	Magnetite (Iron Ore), Fe_3O_4

Of these minerals *feldspars, quartz, pyroxene, hornblende,* and *biotite* are the most important in forming igneous rocks, and consequently the student should make careful note of them. For details regarding them the Appendix that deals with the minerals mentioned in this work should be consulted. It will be seen by examining the chemical formulas given in the table above that the minerals are composed of silica and the six metallic oxides previously mentioned as occurring in the magmas.

Furthermore, since it was shown that the chemical composition of magmas varies, it is evident that the relative quantities of the minerals

in the resulting rocks will also vary. Thus, a siliceous magma, in which $(Na,K)_2O$, Al_2O_3, and SiO_2 are the chief substances, will form a rock that consists mostly of feldspars, whereas a basic magma, in which CaO, MgO, and FeO are high, will make a rock that contains mostly pyroxene, hornblende, and other *ferromagnesian* minerals, as they are called in allusion to the iron and magnesium in them.

Thus it appears that on account of the diverse compositions of magmas the igneous rocks that have formed from them vary both in the kinds and in the relative amounts of their component minerals. These variations in mineral composition largely determine the different varieties of igneous rocks, and mineral composition is one of the principal factors used in classifying the igneous rocks.

Classification of Igneous Rocks

The features by which the igneous rocks may be classified have now been explained. We have seen that the igneous rocks vary in texture and in composition. Both of these variables will be used as factors in building up a classification of the igneous rocks. By employing texture as the principal criterion we at once obtain five major classes: I, even-granular, in which all the minerals are of about the same size and are sufficiently large to be identified by the eye alone or aided by a pocket lens; II, porphyritic-granular, in which certain minerals in virtue of their large size contrast conspicuously with those which surround them, thus forming a porphyry having an even-granular groundmass; III, porphyritic-aphanitic, in which the conspicuous crystals — the phenocrysts — are set in an aphanitic groundmass; IV, aphanitic, in which none of the constituents are distinguishable; and V, glassy, in which few or none of the constituents have crystallized. These five classes are termed *massive* rocks, and for the sake of completeness Class VI is added to take care of the fragmental products of volcanic eruptions. This order from Class I to Class V marks in a general way the decreasing amount of easily recognizable minerals in rocks: in Class I, all the constituents are easily recognizable by the unaided eye; in II, the phenocrysts are easily distinguishable, the constituents of the groundmass less readily; in III, the phenocrysts alone are distinguishable; and in IV, none of the constituents can be recognized.

Each of the major classes is then subdivided on the basis of composition — on the kinds of minerals present and the proportions in which these minerals occur. It is to these subdivisions that the actual rock names are given. For example, an even-granular rock that is composed of feldspar, quartz, and a dark mineral, generally biotite, is called granite.

By applying these principles, the following classification of igneous rocks is obtained, as shown in the subjoined table.

TABLE OF IGNEOUS ROCKS

	Major Classes (based on texture)	Subdivision of Major Classes (based on mineral composition)			
		Light-colored minerals, chiefly feldspar, predominate		Dark minerals predominate	Dark minerals entirely
I	Granular (with grains interlocking)	GRANITE (has quartz)	DIORITE (has no quartz)	GABBRO, DOLERITE (grain size is intermediate between that of gabbro and basalt)	PERIDOTITE, Hornblendite, Pyroxenite
II	Granular (as above) and porphyritic	GRANITE PORPHYRY (has quartz)	DIORITE PORPHYRY (has no quartz)	GABBRO PORPHYRY	
III	Porphyritic, with aphanitic groundmass	RHYOLITE (contains phenocrysts of quartz)	ANDESITE	BASALT	
IV	Nonporphyritic, aphanitic	FELSITE			
V	Glassy	OBSIDIAN Pitchstone Pumice		BASALT GLASS	
VI	Fragmental	Volcanic tuff and breccia			

Note: Syenite is briefly mentioned in the text.

Remarks on the Table. — Leaving out of account the glassy rocks, which are rare, and the tuff and breccia, which are described on page 255, the following remarks may prove of service in understanding the classification of igneous rocks shown in the table.

All rocks in the same horizontal column have the same texture.

All rocks in the same vertical column are essentially of the same chemical composition; for example, granite, granite porphyry, and rhyolite are alike in chemical composition. In physical appearance, however, they differ notably: a granite differs somewhat from a granite porphyry, and vastly from a rhyolite. These differences, as already

pointed out, are mainly the results of the different rates of cooling — the granite has cooled extremely slowly, whereas the rhyolite has chilled rapidly.

The rocks in which the light-colored minerals predominate are light in color and light in weight, *i.e.*, they are of low specific gravity. The rocks in which the dark (ferromagnesian) minerals predominate are dark in color and heavy in weight. The range in specific gravity — from 2.67 for the average granite to 3 for gabbro — is not large, but is sufficient after experience to serve as an aid to identification.

Although in the table each rock has been put in a separate compartment, in nature no rock variety is sharply bounded from its neighbors that are shown in the table. There are for example transitional varieties between granite and diorite and between granite and granite porphyry. No hard and fast boundaries set off any of the so-called rock species. These facts often make it difficult to classify a given rock. It may as well be recognized at the outset that the determination of rocks is at best a difficult matter, and that to classify accurately with the unaided eye the finer-grained and especially the aphanitic rocks, is as a rule impossible. When the accurate identification of a rock becomes a matter of high importance, recourse must be made to the microscope.

Method of Study. — The classification that has just been described is based on what can be recognized by the eye, aided, perhaps, by a pocket lens. It is therefore termed a *field* classification and sometimes *megascopic* (Greek *mega*, great), in contrast to one based on results obtained microscopically, by the study of thin rock slices. Rock slices, or thin sections, are made by cementing a chip of rock to a piece of glass and grinding it down until the section is one-thousandth of an inch thick. In such a thin section, for example, the minute mineral grains that make up the most fine-grained and blackest of basalts become transparent, and can be determined under the microscope (Fig. 196). In this study polarized light is used, and a general knowledge of minerals, of their crystal characters, and optical properties is necessary. It would require too much detail to describe further this mode of studying rocks, which combined with the examination of them by chemical means has developed into a separate geological science, called *Petrology*, the science of rocks. It should be stated, however, that so much additional information has been gained by these methods that the precise classification of igneous rocks is much more complicated than the simple scheme outlined above.

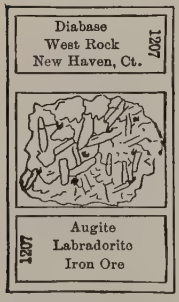

Fig. 196. — Thin section of a rock.

Granite. — As may be seen from the scheme of classification, granite is composed chiefly of quartz and feldspar (of the variety orthoclase). It contains also as a rule a variable amount of flakes of mica, less commonly of hornblende, or both. These component minerals are roughly of the same size, and hence granite is said to be even-granular, or equigranular. As the quartz was the last mineral to separate from the magma, it is molded around the earlier minerals and occupies the angular interspaces between them. This habit of the quartz produces an intimately interpenetrating and interlocking arrangement. This interlocking, even-granular texture is so characteristic of granites that it is often called for short the granitic texture. It serves to distinguish rocks of Class I from all others. The average granite contains 60 per cent of feldspar, 30 per cent of quartz, and 10 per cent of dark minerals. There are many varieties of granite, based on color, texture, etc. Its common occurrence is shown in the fact that there are few states in the Union or provinces in Canada that do not contain exposures of granite; and its use as a building stone and for various other purposes is well known.

Granite is the most important intrusive igneous rock, and appears to be the main constituent of the foundation of the continental masses. These granites are of very ancient origin, of pre-Cambrian age, and constitute a floor upon which the sedimentary rocks of later age were deposited. Granite of younger age occurs also as stocks and vast batholiths that are intrusive into the younger rocks. In all of these occurrences it is either in the form of normal granite, or in a certain modification of it known as granite gneiss. As granite is formed at some depth in the crust, it is exposed at the surface only after prolonged denudation; hence it is seen chiefly in those parts of the continents bared by erosion — that is to say, in mountains or in regions so deeply eroded that the roots of the mountains are visible.

Syenite. — This is like granite in texture and composition but differs in containing little or no quartz. Several varieties are distinguished, based on the character of the feldspar and the accompanying feldspar-like mineral. Thus in *syenite proper* the feldspars are alkalic, that is, contain soda and potash, but little or no lime. In another variety, a feldspar-like mineral, *nephelite* ($NaAlSiO_4$), is present in addition to the feldspars, and the rock is known as *nephelite syenite*. The syenites are not common rocks, nor as a rule do they occur in very large masses compared with granite.

Diorite. — Diorite is a granular igneous rock composed of feldspar and one or more dark minerals, in which the feldspar is more abundant than the dark minerals. The feldspar is mainly plagioclase, but it is generally difficult to recognize this fact with the unaided eye. The

dark minerals may be biotite, hornblende, or pyroxene and they may occur either singly or together.

Gabbro. — Gabbro differs from diorite in that the feldspar is subordinate and the dark minerals predominate. Hornblende, pyroxene, and olivine are the common dark minerals; they occur singly or together; biotite, though present in some gabbros, is distinctly uncommon. Because of the prevalence of dark minerals, gabbros are dark and of high specific gravity. *Dolerite* is a convenient term for the basic rocks that are intermediate in grain size between basalts and gabbros.

Diorites and gabbros, while abundant as intrusive masses, do not commonly occur in extensive batholiths as the granites do. They are more common as smaller stocks, sills, dikes, and sometimes forming the inner part of thick extrusive masses.

Peridotite. — Peridotite is generally composed of a mixture of ferromagnesian minerals, with olivine — peridote, $(Mg,Fe)_2SiO_4$ — predominating. It is not common and usually occurs as minor intrusions, — dikes, sills, or small stocks. It is very interesting and important, however, as being the source of ores of chromium, nickel, and platinum, and of the diamond. It is generally very dark to black and heavy from the large amount of iron-bearing minerals present. The diamonds of South Africa occur in volcanic pipes composed of this rock, and they have been also found in similar intrusions of it in Arkansas.

Pyroxenite, as its name implies, is composed wholly of pyroxene, and *hornblendite* consists entirely of hornblende. They form as a rule bodies of small size; nevertheless, in places, as at the remarkable platinum deposits recently discovered in South Africa, pyroxenite occurs in vast volume.

Granite Porphyry, Diorite Porphyry, etc. — As may be seen by reference to the table of classification, there are various kinds of porphyry, depending on the coarseness of the groundmass and its composition, and on the kinds of minerals that are embedded in it as phenocrysts. Thus if the groundmass is as coarse as it is in granite we may have granite porphyry, or diorite porphyry. Feldspars are the most common phenocrysts; quartz occurs along with the feldspar phenocrysts, chiefly in granite porphyry; and dark to black flakes or prisms of mica, hornblende, or pyroxene occur in many porphyries. The porphyries are a very common class of rocks, occurring chiefly as minor intrusions: in dikes, sills, and laccoliths, and often in necks; they do not occur as batholiths. They also compose many extrusive lava flows. Intrusions of porphyry in the Rocky Mountain region are very common, and in many places are accompanied by valuable deposits of gold, silver, copper, lead, and other ores. Examples of this are seen at Leadville and other places in

Colorado, Montana, Nevada, etc. Porphyries rarely make good building stones, as the masses are generally too much divided by joints, but in places they serve as excellent road material. A porphyry is shown in Fig. 195, B.

Rhyolite. — Rhyolite represents the aphanitic lava form of the magma that at depth consolidates as intrusive granite. It contains phenocrysts of feldspar, quartz, and biotite, and rarely of hornblende. The number of these phenocrysts varies within the widest bounds, so that there is every transition between nonporphyritic and highly porphyritic rhyolite. When the amount of the phenocrysts exceeds 25 per cent of the volume, the rock is by some called a rhyolite porphyry. The colors range from white to gray, pink, red, and purple. Rhyolites or andesites with inconspicuous phenocrysts or with sparse or no phenocrysts are termed *felsite*.

Andesite. — Andesites are of many colors, but in general they are darker than the rhyolites; dark gray is common. On the one hand they are transitional into rhyolites; on the other, into basalt. The average or typical andesite occupies the intermediate position. The darker andesites are of basaltic appearance, but unlike basalts their freshly broken thin edges are translucent when held in bright light. The phenocrysts in andesites commonly consist of striated feldspar and one or more dark minerals (hornblende, pyroxene, or biotite). Quartz phenocrysts are absent (distinction from rhyolite). Andesites that are crowded with prominent phenocrysts are by some called andesite porphyries.

Andesite and andesite porphyry are enormously abundant among the extrusive rocks of the globe. They are the chief products of the volcanoes that form the "circle of fire" surrounding the Pacific Ocean. In fact, it was because of their prevalence in the *Andes* of South America that they were given their name. In virtue of their great abundance and differences in color, texture, and mineral composition the variety of andesitic rocks is legion.

Basalt. — When the color of the lava is very dark gray, dark green, brown, or black the rock is *basalt*, the common effusive equivalent of the ferromagnesian magmas. It may be either compact or vesicular. If the vesicles have become filled with some mineral, such as calcite, chlorite, or quartz, the fillings are called *amygdules* and the rock is termed an *amygdaloidal* basalt. Many basalts are without phenocrysts, but others contain numerous conspicuous phenocrysts, consisting of feldspar, olivine, or augite, or some combination of these. The phenocrysts are hard and have straight, clean-cut boundaries, whereas amygdules are generally soft and have irregular, roundish, or elliptical boundaries. The effusive

occurrence of basalt has been already treated under volcanoes and eruptions. The enormous tracts of land in western America, in India, and elsewhere that were flooded by outflows of basalt have there been mentioned.

Dolerite is the name given to the coarser-grained basalts, in which the grains are so well developed that the constituent minerals are nearly or quite recognizable. There is no hard and fast line between basalt and dolerite on the one hand and dolerite and gabbro on the other.

Felsite. — The difficulty and often impossibility of discriminating between rhyolites and andesites that are devoid of phenocrysts makes it necessary to use an elastic non-committal name. For the light-colored rocks of this class, namely those which are white, light to medium gray, light-pink to dark red, pale yellow or brown, purple or light-green, in short those that are not dark green, dark gray, dark brown, or black, the term felsite is often convenient.

Glassy Rocks. — Volcanic glasses occur as thin crusts on the surface of lava flows or where a lava flow has been very quickly cooled, and they are mostly limited to siliceous magmas. Brilliantly lustrous volcanic glass is called *obsidian*, and the duller and more pitchy variety is called *pitchstone*. *Pumice* is frothed glass. Obsidian was much used in past times by primitive peoples in making weapons, implements, etc. The ancient Mexicans were especially skillful in fashioning knives and razors from it. Natural glasses, like the obsidian of Yellowstone Park, commonly contain crystallized minerals that occur in spherical forms having a radiating or spoke-like structure, known as spherulites.

Obsidians are commonly dark-colored or even black; and yet many of them correspond in chemical composition to rhyolite and granite. Hence they appear to contradict the general rule that nearly all rocks with this composition are light-colored. However, if a piece of black obsidian is chipped to a thin edge it transmits the light and loses much of its dark appearance. The deep coloring is the result of uniform distribution of a relatively small amount of dark material in the glass.

Basalt glass is of rare occurrence. Its formation requires extremely rapid chilling of basaltic magma.

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CHAPTER XII

WARPING, FOLDING, AND FRACTURING IN THE EARTH'S CRUST

The outer shell of the Earth is not fixed and rigid, but undergoes changes that result in movement of some parts with relation to others. Evidence is overwhelming that this has occurred repeatedly in the past in all places where it is possible to examine the structure of the Earth's crust. Movements of the different parts of the outer shell have been not only up and down, but also back and forth in directions parallel to the Earth's circumference. Evidence of such movements is both direct and circumstantial. On many occasions, and even within the present century, there have been abrupt, catastrophic shifts through many feet or even yards, along local fractures that penetrate deeply into the rocky crust. From historic records it is known that gradual movements have taken place, with results that are perceptible only after a long time interval; and slow changes of this kind are in progress at the present time. Back of human history we read the record of crustal movements in obvious deformation of the rocks, ranging from broad, gentle bending or warping to more localized severe folding or fracturing.

All movements of the lithosphere, resulting in relative vertical or horizontal changes of position and in deformation of rocks, are comprehended under the general term *diastrophism.* Movements that affect all or a large part of a continent are termed *epeirogenic,* from the Greek *epeiros,* a continent. More localized disturbances related to mountain building are designated as *orogenic,* from the Greek *oros,* a mountain. These terms are useful in discussion; but a systematic treatment of deformation may well begin with the more obvious effects, of whatever kind, produced either in historic or in late geologic time, and proceed to the results of more ancient movements.

DEFORMATION IN LATE GEOLOGIC OR RECENT TIME

Datum Surface. — The fact and the amount of any recent movement are determined from the relative positions of features on the Earth. Most of these movements are in the vertical direction; and in order to determine the extent and the rate of change it is necessary to have a convenient horizontal surface to which reference can be made. The

average level of the sea is the most logical surface for this purpose, as the shoreline, at mean tide level, is essentially horizontal throughout its whole extent. Any local or differential movement, upward or downward, of land bordering the sea is clearly evident to one who observes the peculiarities of shorelines.

The idea that the sea surface is undistorted and permanently fixed is not strictly correct. Adjacent to high continental borders the water is attracted laterally and upward by the land mass, and the water surface is slightly farther from the center of the Earth in such localities than it is along low, flat coasts or adjacent to oceanic islands. Moreover the sea level has varied within recent geologic times, first through withdrawal of much water from the oceans during the accumulation of continental ice sheets, and later through restoration of this water by melting of the ice. Further, there are reasons for believing that the oceans have increased in size and depth through geologic time by the constant addition of magmatic waters; and it is probable that the ocean basins have changed appreciably in size many times through upward or downward bowing of the floors and by slow filling in of sediment, with consequent raising or lowering of sea level. But such changes are very gradual, and their effects in shifting shorelines are essentially uniform all over the Earth; whereas many movements of the land are relatively rapid, and all such movements vary in amount from one place to another.

Elevation. — The most striking proofs of uplift of the land consist in the locally elevated position of features that we definitely associate with the sea or its edge. Thus in various parts of the world outcrops of rocks with attached shells or skeletons of dead marine organisms, such as barnacles and corals, are found high above sea level. In some localities the rocks are pierced by tubes that were drilled and lined by a peculiar rock-boring marine animal (*Lithodomus*). Excellent evidence of continuous or recurrent elevation is furnished by certain islands of the East Indian Archipelago. In shallow waters off the coast of Timor corals are building extensive reefs. Similar reefs, strikingly fresh but entirely dead, extend from the littoral zone to the higher ground above the reach of the highest tides; and still others, showing various degrees of weathering, occur at different levels up to several hundred feet above sea level. A classic example of changes in land level is found in the temple of Jupiter Serapis built by the Romans near the seashore in the vicinity of Naples. The three columns left standing are bored by lithodomi to a height of 20 feet above the floor, and their shells remain in some of the holes. From this we infer that after the temple was built the land subsided more than 20 feet, carrying the temple into the sea,

and that later there was uplift of about the same amount. This conclusion is confirmed by historic record.

Strong testimony is furnished also by the abnormal position of conspicuous features made by erosion and deposition along a coast. In parts of California, Chile, Scotland, and numerous other coastal regions raised beaches, accompanied by wave-cut and wave-built terraces, form nearly level benches of country terminated inland by former sea cliffs. Typical wave-formed caves pierce the cliffs (Fig. 197), and old stacks, now high and dry, rise abruptly from the terraces. Such an

Fig. 197. — Ancient sea caves in former sea cliff at back of elevated beach, showing strand line. Coast of Fifeshire, Scotland. (Geol. Surv. of Scotland.)

elevated terrace, with its related features, is often spoken of as a raised *strand line*, since it commonly appears as a more or less distinct topographic line, or level, approximately parallel to the present shoreline and above it.

Still another kind of evidence, of a direct and positive character, is furnished by careful observations made year by year. Thus in countries bordering the Baltic Sea an uplift has been under observation for a long period, and has been measured by marks placed on the shores. In some places the elevation has been as much as 3 feet in a century, but the rate is not everywhere uniform and it varies also from time to time. All the facts indicate that the Scandinavian peninsula has risen gradually for a long period, so that the northern part of Sweden is about 900 feet higher than it was at the close of the Ice Age. Raised strand lines

are a noticeable feature in many northern regions (Fig. 198). There is similar proof, on a large scale, that within recent geological time the west coast of South America has experienced very considerable elevation; and probably the uplift is still going on.

Fig. 198. — Elevated strand lines cut in sandstones and limestones. Straits of Belle Isle, Labrador. (Schuchert.)

Depression. — Evidence of subsidence below sea level is less striking than that of elevation, but not less convincing. It is necessary to use care in drawing conclusions, however, because encroachment of the sea upon the land is not in itself a proof of subsidence, as it may result merely from landward erosion by waves and currents. Submergence of features that are definitely characteristic of land surfaces constitutes the best proof. Increasing depth of average water level over well-known rocks or reefs in harbors gives evidence of slow sinking still in progress.

Submerged stumps and other marks of former forests are found at various places along the Atlantic coast from Maine southward. Some of the stumps remain rooted in the old forest soil, covered by marine muds or other deposits (Fig. 199). It is clear that this could occur only in situations, such as protected nooks and corners of estuaries, sheltered from the waves of the encroaching sea which would otherwise have swept away the forest soil. Submerged peat bogs, made under fresh-water conditions but now turned into tidal flats, are rather common and tell a similar story. All the cumulative evidence goes to show that the Atlantic seacoast from Maine southward has gradually sunk within a late geologic period. Whether parts of it are still sinking is a matter about which geologists have not yet reached agreement.

Drowned Valleys. — Evidence of subsidence of the land is furnished by the irregular shorelines produced by the drowning of valleys, with production of bays and estuaries. The seaward extension of river channels, such as the Hudson, for long distances across the submerged continental shelf, demands the same explanation; for manifestly these great trenches, now sunk in the sea floor, could not have been cut while the continental shelf was covered with water, but only by river or glacial action, or both, when it stood at a higher level and was a land surface. The submerged channel of the Hudson has been outlined by closely spaced soundings on the continental platform south of New York harbor. It has the definite form of a valley, extending more than

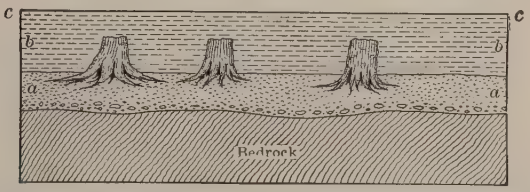

Fig. 199. — Showing submerged forest. *a*, old forest soil with stumps standing in it; *b*, marine deposits of silts and sands; *c–c*, present level of high tide.

100 miles southward from Sandy Hook to the steep continental slope; and its depth below present sea level ranges from about 100 feet at the north to more than 2000 feet at the south.

Subsidence and Deposit of Sediment. — In many parts of the world thick deposits of sediment are being laid down by rivers in subsiding basins adjacent to coasts. Borings into deltas pass through alternating marine and fresh-water deposits, or even through sediments entirely nonmarine, to a great depth below present sea level. For example, wells sunk into the delta of the Po, near Venice, pass through four separate layers that contain abundant remains of plants similar to those now growing in the marshlands along the Adriatic. One of the layers is about 300 feet below sea level. At shallower depths some of the sands, gravels, and clays contain shells of marine molluscs, and other layers yield fresh-water types, such as land snails. In the delta of the Ganges, near Calcutta, pieces of wood and bones of land animals are found hundreds of feet below sea level. Deep wells in the Great Valley of California furnish similar evidence. Facts of this kind show that subsidence has been going on for a long period, not at an even rate, but as an interrupted process, whose variations permitted alternating fresh-water and marine deposits to be formed.

Evidences of Elevation or Depression Inland. — Movements of the lithosphere involving changes of level are not confined to the sea coasts;

they occur also in the interior of the continents. For example, in 1811 large areas of the Mississippi flood plain near New Madrid, Missouri, sank far below their former level and now are occupied by lakes. Trees that grew on the plain were killed by the flooding, and for a long time their dead trunks or tops, projecting above the water, furnished evidence of the changes in level.

An excellent illustration of tilting on a large scale is afforded by the Great Lakes. To the northeast the land has risen since the retreat of the great ice sheet, and as a result the lake basins have been tilted southwestward. The old strand lines are several hundred feet above the present water level on the north and northeast, and slope toward it as they are followed west and south. Since the lakes discharge to the

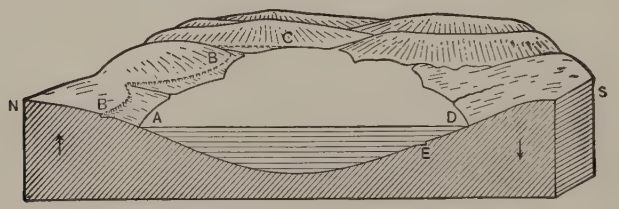

Fig. 200. — Showing tilting of lake basin. *AD*, present lake level. *BB*, raised shoreline disappearing under lake at *C*. Land has been raised on the north (N), and on the opposite side the former shoreline, *C* to *E*, has been drowned.

east, the raising of their outlets has caused them to enlarge, expanding them to the west and south. As a result, river mouths on the south and west sides of some of the lakes (especially Erie and Superior) have been drowned. The tilting movement is still in progress and has been accurately determined. It is at the rate of 6 inches per hundred miles per century. Small as this rate seems, in 1600 years it would cause the upper Great Lakes to discharge by way of Chicago River into the Mississippi drainage (Fig. 200).

Significance of Recent Changes in Level. — It is obvious that changes of surface level must be accompanied by bending or fracturing of the lithosphere. In fact, surface movements merely record deep-seated processes which lead to slight changes in the form of the globe. Commonly the movement is a broad warping that affects wide areas. Locally there may be sharper bending, or the rocks may actually break under the enormous strain, causing abrupt offsetting of roads or other features at the surface. Such breaks, or *faults*, are common along the Pacific coast of North America, in Japan, and in many other parts of the world. Displacement along some of these fractures within historic time has caused local shifts in level or horizontal changes of position.

In considering effects of this kind we are impelled to inquire at once into their ultimate cause. Discussion of this question will be postponed until more of the essential facts and conditions have been presented. In general, however, it does not seem strange that a globe so large as our Earth, which spins rapidly on its axis and whirls through space, which suffers chemical and physical changes in the outer and probably also in its deeper parts, should be subject to slow and almost continuous deformation. Volcanoes testify to local unstable conditions, which result in the melting and moving of rock material. Erosion and sedimentation through long periods result in the transfer of great loads at the surface; and this process undoubtedly sets up enormous strains in the lithosphere. These well-known processes would in themselves cause some slow deformation of the crust; and probably more profound changes in progress in the deep, unknown parts of the Earth are responsible for still greater deforming forces.

Regardless of ultimate causes, it is desirable that we get from the actual facts a picture of the Earth as a changing, dynamic thing. "Terra firma" is not literally fixed and inert. Parts of the Earth's crust are changing in form and position today, and all parts have moved at some time during geologic history.

RESULTS OF ANCIENT CRUSTAL MOVEMENTS

It is neither desirable nor possible to distinguish sharply between recent and older deformation. Certain features at the surface bear unmistakable witness to very recent crustal disturbance, although there may be no direct record of the event in human history. On the other hand we see in the rocks many fractures and folds that date from very early periods in Earth history, as is shown clearly by geologic relationships. Between these two extremes are found indications of disturbance in every geologic epoch, showing that movements of the same or similar kinds have been continuous or recurrent throughout recorded geologic time. In general the record of last movement is found in forms and features on the Earth's surface, such as the elevated beaches, drowned valleys, and tilted lake basins described above. All surface forms are ephemeral, because of erosion; and as they disappear or become fragmentary, the most reliable guide to former crustal movements is found in the structure of the underlying rocks.

Study of the framework or structure of the lithosphere is called *structural geology.* Such a study may be entirely geometrical, with the object of furnishing an exact description of the crust, including all breaks and bends in the rock units. Knowledge of this kind has great economic

value, as it is essential in locating and following beds of coal or mineral veins. However, the structural features of the Earth have also a broader interest and value, as they furnish the clue to important events in the history of a region.

The principal structural features that indicate deformation of the lithosphere are broad bends or *warps*, *folds* of various sizes and degrees of intensity, and *fractures*.

Broad Warps

Certain topographic features that result from gentle warping of the crust develop slowly and persist for a very long time. For example, when a peneplain is domed up widely the streams become deeply entrenched and their old meander patterns are preserved in the deepened valleys. As the deepest entrenching results in the area of greatest uplift, the middle portion and the edges of such a great dome can be recognized by study of the entrenched valleys. Furthermore remnants of the uplifted peneplain surface will persist for a long time between the valleys. By study of the valleys and the remains of old surfaces it is determined that the Appalachian region is an irregularly domed peneplain undergoing dissection. It is obvious that evidence of this kind, used for recognizing warping movements that are quite old, is similar to that by which recent uplift is determined.

A more permanent record of gentle warping is found in the bending or tilting of stratified rock formations that originally were nearly or quite horizontal. In southeastern Texas marine limestones and shales deposited in Cretaceous time now lie slightly above sea level, with a gentle inclination toward the Gulf. Formations of the same age, containing fossils of identical marine forms, are exposed in eastern Colorado, 5000 feet and more above sea level. When these beds were laid down a continuous seaway reached northwestward from the Gulf of Mexico across the United States. In later time this part of the continent was elevated by a broad warping movement, so that the formations in Colorado were lifted a mile higher than the corresponding beds in Texas, 700 miles away. In the Colorado Plateau of Utah and Arizona a thick blanket of marine strata lies thousands of feet above sea level, and is dissected by deep canyons. In a general way these strata are nearly horizontal; but if any one layer is followed and mapped in detail it is found to bow gently upward into irregular domes and bend downward into shallow basins. It is clear that this wide area of sedimentary beds, many hundreds of miles across, was not lifted up with absolute uniformity, but was bent somewhat, so that each layer now resembles a wide board that has been warped by exposure to the weather. The original

surface of the uplifted mass has been entirely destroyed by erosion; but a record of the distortion is preserved in the form of each layer of stratified rock beneath that former surface.

The value of widespread marine sediments as a means of detecting such gentle warping of the crust is obvious. In areas occupied entirely by granites and similar massive rocks, a broad warp is recorded only in the surface forms; and after these forms have been obliterated by erosion there is nothing to suggest the ancient movement.

Folds

Anticline and Syncline. — In many places the stratified rocks have been buckled into plications or folds. Some of these are on a small scale and can be seen directly; but commonly the folding is on such a great scale, and exposures of the rocks are so discontinuous, that it is necessary to study and piece together the structure of certain distinguishable layers over many miles before the form of the folds becomes clear. The nature and scale of the folds in mountain regions can be appreciated best by study of maps and cross sections that have been prepared by geologists in the field.

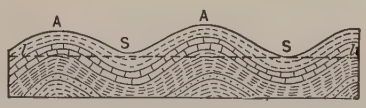

Fig. 201. — To illustrate anticlines, *A*, and synclines, *S*.

Two terms are used constantly in descriptions and discussions of folds. Like regular swells on the ocean, rock folds ordinarily occur in a series, with alternating crests and troughs. The crests of the folds — that is, the upfolds — are *anticlines;* the troughs, or downfolds, are *synclines* (Fig. 201). Even if the original surface crests should be carried away by erosion, and the whole reduced to a level plain, we should still call the upfolded portions below the surface anticlines, the downfolded

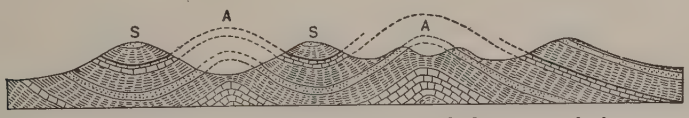

Fig. 202. — Anticlines, *A*, and synclines, *S*. Folds have no relation to present surface forms.

portions synclines, and in imagination reconstruct the missing parts (Fig. 202). Thus it should be clear that anticlines and synclines *are not a matter of surface topography, but of structure* (Figs. 203 and 204). Commonly the original configuration of the surface is even reversed by erosion, so that valleys now occupy the positions of the ancient crests, and ridges or mountains are in the places of the troughs; but the original structural terms still apply (Fig. 202).

Fig. 203. — An anticline broken at the top. In the foreground the outcrops of the eroded strata are seen dipping outward, from which the anticline structure could be inferred if the arch did not exist. Pembroke, Wales. (Geol. Surv. of England and Wales.)

Fig. 204. — A syncline, near Hancock, Md. (U. S. Geol. Surv.)

Outcrop. — Only the ideal relation of simple, upright, regular folds has been considered above. A series of folds approximating this form is by no means uncommon in nature; but usually the folding is much more complicated. Moreover some important facts about folds cannot be represented in ordinary cross sections. The varied kinds of deformation which the rocks have suffered in any region condition the *geologic structure* of that region; and it is a matter of the highest importance that the geologic structure of every country should be known so far as possible and represented accurately on maps. If the surface of the Earth were everywhere naked bedrock, this would be relatively an easy matter; but since the rocks have been greatly eroded, and are largely covered with earth and vegetation, or with water, snow, and ice, the natural difficulties of the task are enormous. The structure in a region is determined by a careful study and comparison of the *outcrops*, by which term is meant the actual exposures of bedrock at the surface. If the ground were perfectly level and the strata horizontal, the outcrop would be the flat surface of the uppermost rock stratum, and we should learn little from it; but on slopes and cliffs bordering stream valleys, we may inspect the outcropping edges of many strata. If the sides of the valley were trenched by ravines, the line of outcrop would not be straight, but sinuous, retreating from the valley into the ravines, and advancing on the spurs (Fig. 205).

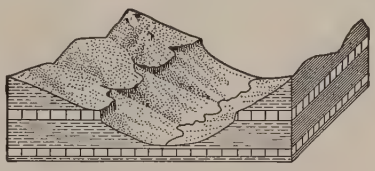

Fig. 205. — Section and outcrop of horizontal strata along valley.

If the strata have been inclined by folding and eroded their edges may be exposed even on a nearly flat land surface. Commonly the edges of the harder, more resistant beds project to form the more prominent outcrops. In mountain regions, soil and other concealing débris usually decrease in amount with increasing height; and exposures of rock grow in prominence correspondingly, until the upper rocky ridges and peaks may each be a vast outcrop. Because of the excellent exposures, and the great depth of the section visible in canyon walls and on the cliffy slopes, mountains furnish the most favorable opportunities for determining geologic structure.

Dip and Strike. — These terms, used constantly in describing the attitude of inclined strata, may be defined as follows: *Dip is the angle of inclination of the plane of bedding from a horizontal plane.* *Strike is the direction of the line of intersection of the plane of bedding with a horizontal plane.* Reference to a diagram will make the definitions clear. In Fig. 206 any horizontal line, as AB, drawn on the surface of the in-

clined stratum represents the strike. Lines with any other position on this surface, as *EF* or *GH*, are inclined; but *CD*, at right angles to

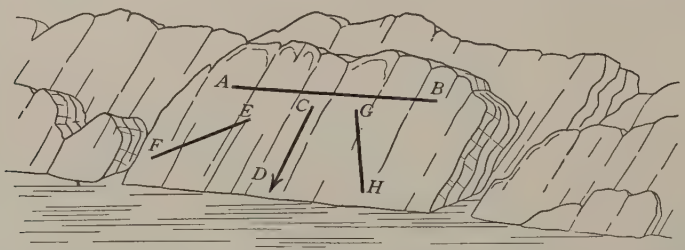

Fig. 206. — To illustrate strike and dip of tilted strata. The line of intersection of the horizontal water surface with a bedding plane gives the strike; or any other horizontal line on the bedding, as *AB*, also represents the strike. The lines *EF* and *GH* have an inclination, but the line *CD*, which is at right angles to *AB*, shows the maximum or true dip of the layer.

the strike line, represents the maximum inclination of the surface, or the *dip*. The practical use of strike and dip are better illustrated in Fig. 207. A dark stratum, *AG*, is exposed on a perfectly flat surface.

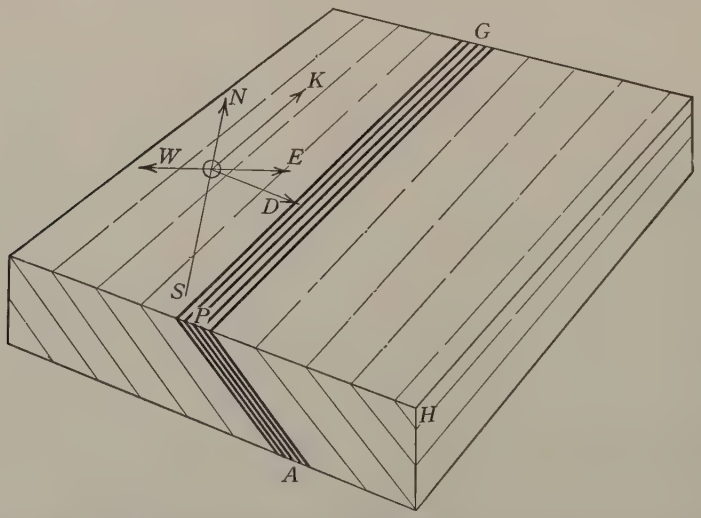

Fig. 207. — To explain strike of strata, and direction and amount of dip. The stratum, *A–G*, is exposed on a horizontal surface. The direction of the strike, with relation to the compass points NESW, is shown by the angle *NOK*, which is east of north. The stratum dips in the direction *OD*, which is east of south by the amount of the angle *SOD*. The angle *APH*, measured from the horizontal plane, is the angle of dip.

The direction of its outcrop, *PG*, is the strike. If this direction, measured by the angle *NSG* (= *NOK*), is 30 degrees east of north, the stratum is said to strike N. 30° E. Obviously a layer with this strike might dip

either toward the northwest or toward the southeast. The stratum in the diagram inclines southeastward, making an angle *APH* with the horizontal. If this angle measures 40 degrees, the stratum is said to dip 40° southeastward; or more precisely, as the direction of dip is at right angles to the strike, and the latter is N. 30° E., we say the dip is 40° S. 60° E. In the diagram, the direction of the dip (S. 60° E.) is shown by the angle *SOD*. The meaning of the statement is clear if it is kept in mind that the first angle — 40° — is the *amount* of the dip, measured downward from the horizontal (angle *APH*, Fig. 207); whereas the second angle — 60° — gives the *direction* in which the stratum dips, with relation to the north–south line.

The direction of strike is taken with an ordinary compass, and the direction of dip is calculated. The *amount* of dip is taken with a clinometer, which is essentially a pendulum swinging over a graduated arc (Fig. 208). For geologic purposes the compass and clinometer are usually combined in one instrument.

Dip and strike are represented on geologic maps by a conventional sign $\overline{\downarrow}$, in which the direction of the cross bar, as placed on the map, indicates the direction of strike, and the arrow points in the direction of

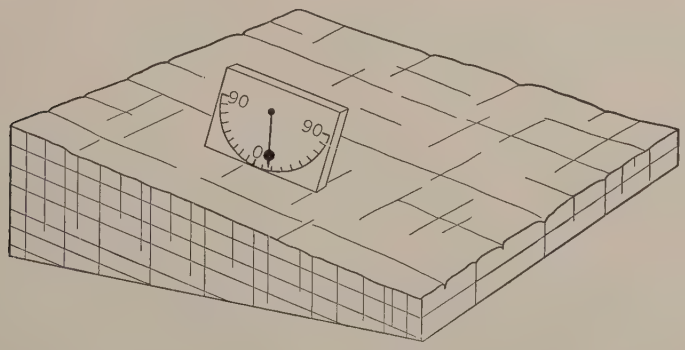

Fig. 208. — To explain measurement of dip angle with the clinometer. The pendulum of the instrument swings freely on an axis, and therefore is always vertical when the box is on edge. When the edge of the box rests on a bedding plane in the direction of dip, the angle of dip is read directly on the graduated arc.

dip (Fig. 209). The length of the arrow is also sometimes used to show in a general way the amount of dip; thus $|\!\rightarrow$ indicates a low angle of inclination; $|\!\rightarrow$ a steep dip. Ordinarily the actual amount in degrees is written in; *e.g.*, $|\!\rightarrow$ 30°.

Pitching Folds. — A series of dips and strikes arranged on a map as shown in Fig. 209 indicates a set of parallel folds in the underlying rock. It is quite evident that folds of this kind could not run in the direction

of strike indefinitely, or around the world; they must end somewhere. The end of a syncline, as seen on a map and in section, has the appear

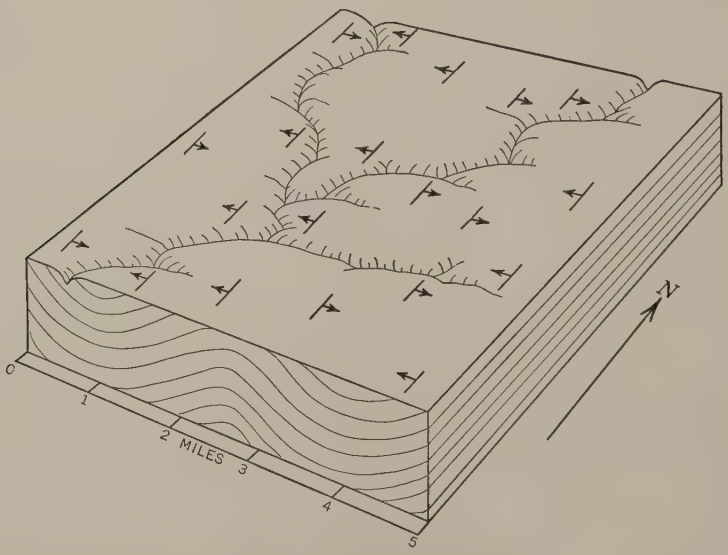

Fig. 209. — Diagram of a land surface underlain by folded strata. In making a geologic map of this area, the strike and dip of each outcrop is noted in the proper position, by use of the conventional symbol ⊤. The symbols in the figure indicate the positions of an anticline and two synclines, which are shown in section at the end of the block.

ance shown in Fig. 210, A. A stratum is warped into a form like the end of a boat. An anticline, near its termination, resembles a boat overturned (Fig. 210, B). In either of these structural features the outcrop

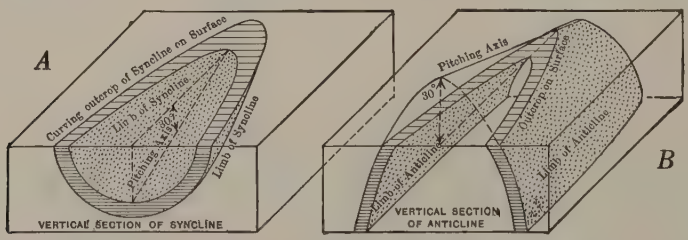

Fig. 210. — *A*, ending of a syncline, as seen on a flat surface. *B*, near the ending of an anticline. For simplicity, only a single stratum is shown in each case. Consider that each block is about half a mile wide. (H. H. Robinson.)

of a stratum, on a nearly flat surface of erosion, turns in the shape of a horseshoe; but there are two general ways for distinguishing one form of fold from the other. In the ordinary syncline, dips are consistently toward the inside of the horseshoe (Fig. 211, A), and the younger strata

lie inside. In an ordinary anticline, dips are outward at every point, and the older strata lie inside the horseshoe (Fig. 211, B).

The median line of a fold, along the top of an anticline or the bottom of a syncline, is the *axis*. This line extends along a bedding surface, or along this surface restored if it has been partly eroded (Fig. 210). The axis *emerges* from the ground toward the end of a syncline, *plunges* into the ground toward the end of an anticline. The angle between the axis and the horizontal is called the *pitch* of the fold. It is evident that the pitch is merely a special case of dip measured along the axis. In a

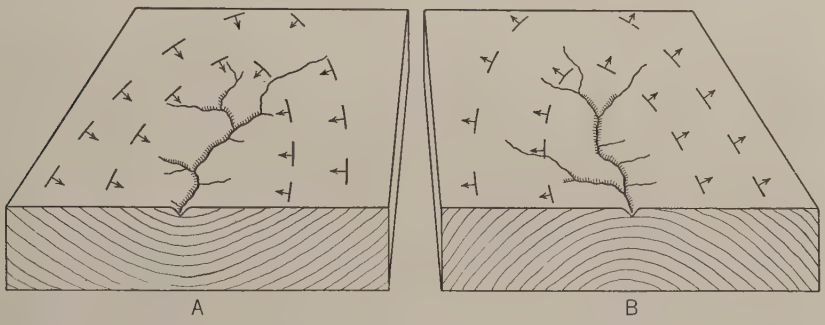

Fig. 211. — Illustrating the use of dip and strike symbols on a map to show a plunging syncline (*A*) and a plunging anticline (*B*).

pitching or *plunging* fold, the sides or *limbs*, as exposed on a nearly level surface, cannot be parallel, but necessarily converge or diverge (Figs. 210–212). If outcrops of strata in the two limbs run along parallel to each other for a long distance, the axis of the fold must be essentially horizontal. In the Appalachian region, hard sandstones or other resistant formations in the limbs of eroded folds make some of the high mountain ridges. Commonly two such ridges, on opposite sides of a great fold, maintain a straight parallel course for many miles; but eventually they converge and unite at the end of the fold.

Inclined, Asymmetric, and Broken Folds. — Thus far we have considered only simple, regular, upright folds. If a plane is imagined to pass through the center of a fold and its axis, as in Fig. 213, like the extended keel of a boat, we may call this the *axial plane* of the fold. In a regular or symmetric fold, this plane is one of symmetry; that is, the parts to left and right of it are symmetrically disposed, or each point on the left of the plane has its corresponding point at an equal distance on the right of it. If the fold is upright the plane is vertical (Fig. 213). However, some folds are not upright but have been pushed over until the axial planes are inclined. A fold of this kind is said to be *over-*

turned (Fig. 214). Such overturning may, indeed, go so far that the axial plane is nearly, or actually, horizontal; and the fold is then termed *recumbent.* Synclines may have a similar attitude.

Fig. 212. — On a plain of marine erosion the outcropping edges of the strata are seen at the ending of a syncline, as shown by the curving strike and inward dip. Near North Berwick, Scotland. (Geol. Surv. of Scotland.)

It is also common to find that folds are asymmetric (without symmetry); that is, they are not similar to right and left of the axial plane,

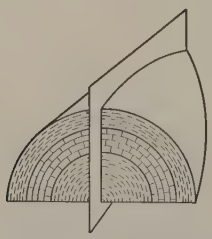

Fig. 213. — Upright symmetrical fold; axial plane vertical.

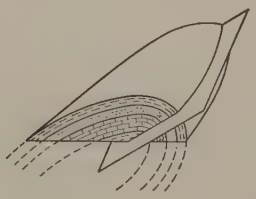

Fig. 214. — Inclined symmetrical fold; axial plane inclined.

which is not, therefore, one of symmetry, as in a regular fold (Fig. 215). Such asymmetric folds may be upright, overturned, or recumbent.

Finally, folds may be so sharply flexed (creased) that they may break, especially at the apex; and on breaking, the parts are likely to be dis-

placed with respect to one another, or faulted. Faulting, however, is so important a phenomenon that it deserves especial consideration in a later place.

Other Features of Folds. — If folds are so sharply flexed that the limbs are nearly or quite parallel, they are said to be *closed;* in this condition the horizontal distance across the strata, or the width of the fold, cannot be farther reduced without squeezing or mashing of the

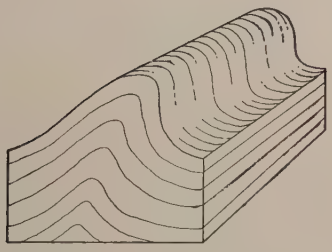

Fig. 215. — An asymmetric fold.

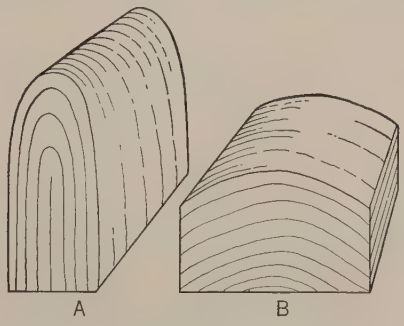

Fig. 216. — *A*, closed fold; *B*, part of an open fold.

beds (A, Fig. 216). If the limbs make a large angle with each other (as in B, Fig. 216) the fold is *open* and the strata may be further folded without mashing.

In *isoclinal* (equal inclination) folds the strata are compressed until, on both sides of a fold, and perhaps throughout a series, they are parallel

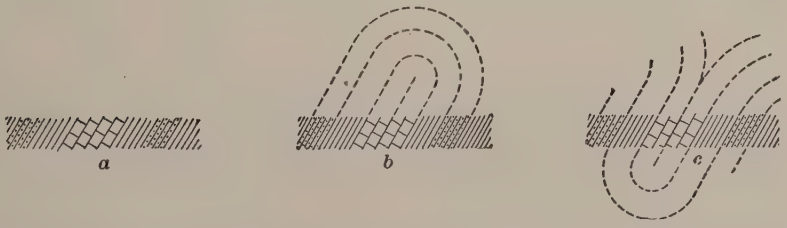

Fig. 217. — Outcrop of strata show in cross section as in *a*; they might be one series with inclined dip, or possibly are in a closed isoclinal fold. Assuming that the strata are arranged symmetrically on opposite sides of a middle line, the structure is an anticline, as indicated in *b*, if the strata are progressively older toward the middle; or it is a syncline, as shown in *c*, if the strata are progressively younger toward the middle.

and have the same dip (Fig. 217, *b* and *c*). When such folds are cut away by erosion as in *a*, some skill is required for correct interpretation of the structure. The term *homocline* is sometimes used to describe a series of bedded rocks all dipping in the same direction. The strata may be

folded isoclinally, or simply tilted uniformly. A special kind of homocline is the *monocline*, in which the strata are bent in one direction only (Fig. 218). A true monocline is a one-limb flexure, on either side of which the strata are horizontal or have uniform gentle slopes.

Folds have been treated thus far as simple structures with true axial planes; but many folds are warped or bent, so that they do not have axial *planes* in the true sense of the word, as the surface bisecting each

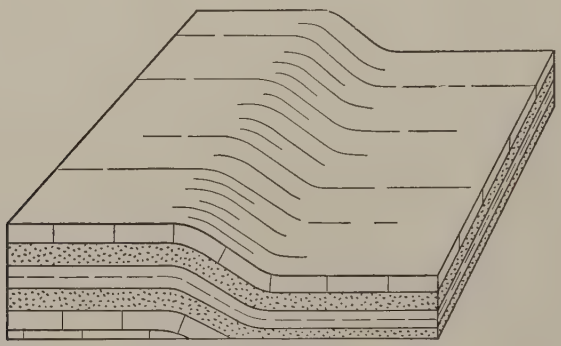

Fig. 218. — A monoclinal fold.

fold is not plane but curved into a sinuous form. Folds may also branch into compound, complicated structures.

Geosynclines and Geanticlines. — Long belts within a continent or on the ocean floor have been warped down to form *geosynclines*, whose dimensions are measured in hundreds of miles. Correspondingly great upwarps are called *geanticlines*. The prefix in each case (from the Greek

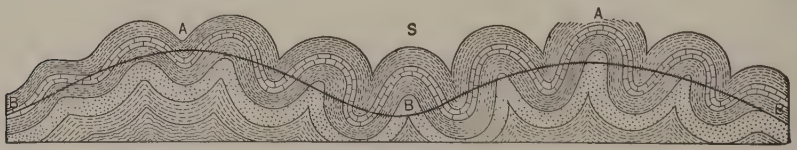

Fig. 219. — Illustration of terms used in compound folding. The general uplifted masses of folds *AA* are called anticlinoria, while the downwarped mass of folds *S* is termed a synclinorium. The general average warping effect of the folding is indicated by the line *BB*.

word *Geos*, meaning Earth) emphasizes the scale of these features. In contrast with ordinary folds, the flexures responsible for geosynclines and geanticlines are very gentle. The Baltic Sea may be a modern geosyncline. In the region about Cincinnati, Ohio, the Paleozoic strata are bowed gently into a geanticlinal arch 250 miles wide.

The geosynclines of the past, as well as those of the present, have been

the great basins for the accumulation of sediments, like those exposed in the Appalachians and the Alps. When later the accumulated beds are compressed into folds, the whole series may form a compound uplifted mass, which erosion carves into mountains. Such a mass of strata, laid down in a geosyncline and crushed into folds, has been termed by Dana a *synclinorium* (from syncline and *oros*, Greek for mountain). The term thus introduced by Dana has, however, been diverted from its original meaning, and applied to a general syncline compounded of minor folds and contrasted with *anticlinorium* (Fig. 219). It has thus become a term of structure, and the related idea of mountain making, which the name expresses, has been relegated to a subordinate position, or entirely left out.

Joints and Faults

In the outer shell of the Earth the rocks are traversed in all directions by fractures, varying from minute crevices to important fissures. We have considered the importance of fractures in the weathering of rocks and formation of soil; in the holding and in the circulation of ground water; and we shall discuss them again in connection with mineral veins. They are, indeed, of great geologic importance, because of the processes that give rise to them and the results achieved by their aid.

A fracture on which there has been no appreciable displacement is called a *joint.* Ordinarily joints are closed so tightly that little or no space is visible between the walls. If the walls are distinctly separated, the term *fissure* is preferable to joint. Some fissures are open, and others have been filled with mineral matter deposited by circulating water. If there has been relative displacement of the walls in a direction parallel to the fracture, so that corresponding points on the two sides are distinctly offset, the fracture is known as a *fault.* Faults are very important geologic features.

Joints in Stratified Rocks. — Field examination shows that joints are common, but that they are much more numerous in some places than in others. Where they are abundant, commonly they are arranged in more or less definite *sets;* that is, the divisional planes running through the rock fall into groups according to direction. In many places there are two prominent sets of joints, approximately at right angles to each other and each set nearly vertical. Such a combination of two or more intersecting sets constitutes a joint *system.* Combined with natural divisional bedding planes, a well-defined system of joints divides stratified rocks into series of closely fitted blocks. The finer the grain of the rock, as a rule, the more perfect the jointing and the more definite the

resulting blocks. Thus, in shale beds and in limestones, the jointing may be very perfect, as illustrated in Fig. 220.

Such jointing may result from various forces; for example, from the tension produced in the beds of sediments by contraction when they are elevated from the sea-bottom to form land masses, and undergo a drying-out process. A more common probable cause of regular fracturing is the warping and twisting suffered by the strata during crustal move-

Fig. 220. — Illustrating joints in limestone beds. The horizontal lines are bedding planes. There are two sets of joints, nearly vertical and at right angles to each other. Drummond Island, Mich. (U. S. Geol. Surv.)

ments. The exact cause of most joints in stratified rocks is not surely known; but they have sometimes been classified as *tensional* or *compressional*, according to the supposed nature of the force producing them. In regions where the strata have been definitely folded, as in many mountain zones, unquestionably the compressive force that plicated the beds produced many joints. More locally, and especially at the crests of anticlines, joints may have been made by stretching or tension.

Strike joints are parallel to the strike of the beds, or nearly so; *dip joints* are essentially in line with the dip, and therefore at right angles to strike joints. *Oblique joints* have intermediate directions. Certain joints extend for long distances across a thick series of beds, and are known as *master joints*. They are contrasted with minor fractures, which may be limited to a single stratum.

Joints in Igneous Rocks. — A common type of jointing in igneous rocks is due to the contraction resulting from the cooling of the original

magma. This occurs during and just after solidification from the liquid state. It may manifest itself in one of several ways, depending on the rate of cooling, the size and shape of the igneous body, and other factors. Thus intrusive masses of granite and similar rocks are characteristically cut by joint planes that divide them into large blocks or prisms. Finer grained masses in sheets, laccoliths, and dikes may be divided into small angular fragments by closely spaced joints. In some laccoliths and similar dome-shaped intrusions there is a shelly jointing on a large scale,

Fig. 221. — "Devil's Post-pile"; columnar jointing in lava. Head of San Joaquin River, Calif. (U. S. Geol. Surv.)

parallel to the domed surface. This appears to have been caused by nearly uniform cooling of the mass from the periphery, with resulting separation into sheets.

The most striking kind of contraction jointing in an igneous rock results in development of *columnar structure*. This result is produced, in general, when two dimensions of the mass are great and the third is small, as in a dike, a sill, or a lava flow. The rock-body may then be composed of a series of closely fitted prisms, which are subdivided by inconspicuous cross joints. The prisms have a variable number of sides, but commonly they tend to be hexagonal, and some of them have remarkable regularity of form (Fig. 221). They may range from several inches to a number of feet in diameter, and up to 200 feet or even more in length. The Giant's Causeway on the north coast of Ireland is one of the most celebrated examples of this columnar structure. The col-

umns form at right angles to the chief cooling surface, and consequently in a level intruded sheet or flow of lava they stand vertically, whereas in a vertical dike they tend to be horizontal. Thus some dikes, exposed as walls by erosion, resemble regularly piled cordwood. In other igneous bodies the position and form of the columns depend on the directions taken by the periphery of each individual mass. Columns in volcanic necks may be vertical, horizontal, or curved; and in some masses they may be arranged radially, like a great fan.

The reason for columnar structure appears to be that in a cooling mass, centers of contraction tend to occur on the cooling surface at equally spaced intervals. From each center three cracks form and radiate outward at angles of 120°. Intersection of these cracks produces a regular hexagonal pattern, and their penetration inward makes the columns. But nature is complex, and the ideal pattern is commonly modified by the occurrence of five- and four-sided figures of varying dimensions. By contracting lengthwise the individual columns may break into sections. The same principles of contraction result in the polygonal shapes commonly seen on mud flats that have cracked from drying (Fig. 155).

In addition to the joints caused by cooling, later fractures caused by crustal movements may affect igneous rocks; but wherever a prominent columnar structure or other well defined fracture system is original in the rock mass, later stresses are more likely to be relieved along these existing breaks than to form additional fractures.

Jointing in Metamorphic Rocks. — As a rule the metamorphic rocks are much jointed. This might be expected, because of the extensive deformation to which such rocks have been subjected. The character of the jointing varies considerably with the nature of the rock. Many of the massive gneisses have joint systems like those characteristic of granite; whereas the fissile and schistose rocks, such as slates, have joints more like those found in sedimentary rocks. (See Chapter XIV.)

Practical Importance of Joints. — Joints are a matter of great importance in all quarrying, tunneling, and mining operations where rock work enters as an important factor, since the jointing obviously facilitates progress. Without them, every rock fragment would have to be broken or blasted loose from bedrock. However, joints may also be a serious inconvenience, especially if large blocks of quarried stone are desired. Perfect monoliths 50 or 100 feet in length can be obtained from comparatively few localities.

General Features of Faults. — Displacement of rock masses along a fracture may occur at the time of the break, or at some later time. Thus a joint might eventually become a fault. Faults are common

features in rocks of all kinds. They are most evident in stratified formations, as the offsetting of layers makes the break conspicuous and directly measurable. However, massive igneous rocks may be faulted as well; and as mineral veins or other features of economic value may be displaced by such fractures, it is important from a practical as well as a scientific standpoint that the nature of faults be well understood.

The surface of fracture along which movement and dislocation has occurred is often spoken of as the fault plane. Although a limited part of it may be nearly plane, it is rarely flat for any considerable distance,

Fig. 222. — Part of an old fault surface, with slickensides, uncovered by erosion. Note the lined and fluted character, indicating that the movement was directly down the dip of the surface. Spotted Range, Nevada. (Longwell.)

but more or less curved, broken, and offset. Therefore it is better, and causes less misapprehension, to term it the *fault surface*. Moreover, the movement in faulting may occur, not upon one surface, but upon a number of more or less closely adjacent breaks, producing a *fault zone*, in which the various offsets make in the aggregate the total displacement. Such a distribution is sometimes called *step faulting*. The masses of rock involved in fault movements are generally of such size and weight, and so compressed together, that the motion of one fault face on the other takes place under tremendous pressure. As a result of the friction, the rock faces are smoothed and striated, and not uncommonly receive a high polish. Such polished and grooved surfaces are known as *slickensides* (Fig. 222). The line of intersection of the fault with the plane of the horizon is called the *strike*, or trend, of the fault, just as we speak of the strike of upturned strata. The surface of faulting is rarely ex-

actly vertical; commonly it is inclined, and in some important faults it approaches horizontality. The angle between the fault surface and the horizontal plane is the *dip*. In an inclined fault the side that overhangs is known as the *hanging wall*, the other as the *foot wall* (Fig. 223). If one were to descend along a fault, as in an inclined shaft of a mine, the appropriateness of these old mining terms would be evident.

Generally the fracture is closed tightly; but parts of it may have been open at one time, and have been filled with mineral matter deposited from solution. Along many faults the grinding of the walls upon one

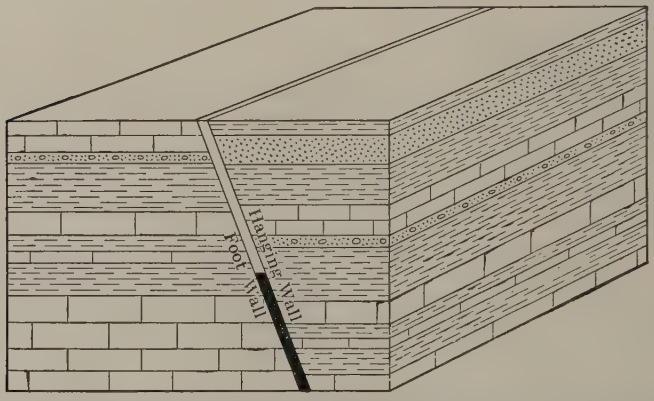

Fig. 223. — To explain fault relations and terms. The strata have been displaced by a fault, and a vein of mineral (black) has formed along the fracture. Mining operations have removed the mineral to a considerable depth, exposing the hanging wall and foot wall of the fault.

another has produced a zone of broken and crushed rock known as *fault breccia*. Commonly there is a thin seam of clay-like material, known as *gouge*, directly along the fault. In the displacement of stratified rocks the friction usually causes bending of the layers near the fault surface. This feature, referred to as *drag*, may be a useful aid in determining the relative direction of motion on the two sides of the fault (Fig. 224).

The features explained so far have to do chiefly with faults as seen below the surface of the ground. Ordinarily a fault breaks the surface as well as the rocks beneath; and if one side of the fault is elevated with relation to the other, the result is an abrupt cliff, or *fault scarp*. With the passage of time the original scarp is modified or even removed by erosion.

Motion on the Fault. — If we assume that one side of a fault stands fast, motion on the other side may be vertical, horizontal, or oblique.

Thus in Fig. 225 the lettered plane may represent one fault face, which for convenience is considered to remain at rest. It is exposed by removal of the block which is assumed to have moved. If we suppose that

Fig. 224. — Fault in shale; the drag of the beds shows that the left side has gone down, the right up. Little River Gap, Tenn. (U. S. Geol. Surv.)

some particle at A, for example a crystal, was cleft by the fault, then one part A remained in its original place and the other part, embedded in the opposite face, was carried in some direction by the faulting. Suppose that the moving portion of the crystal scratched a groove on the stationary fault face. Depending on the direction of movement, this scratch or *striation* might be one of the lines AH, AG, AC, or AD. Most commonly it would be in some oblique direction, as AB.

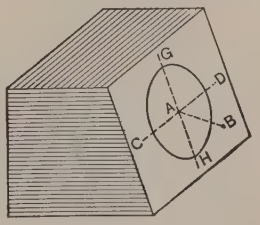

Fig. 225. — To show possible motion in faulting. The hanging-wall block is omitted, to show entire fault surface.

Normal and Reverse Faults. — If faulting takes place by movement upward or downward, two different kinds of structure may result. In A, Fig. 226, the hanging wall has apparently slipped down with reference to the foot wall; a fault of this kind is known as a *normal* fault. In B of this figure, the hanging wall has apparently been crowded up over the foot wall; a fault of this kind is called a *reverse* fault. In the normal movement a particular layer V–V has

been lengthened apparently by an amount corresponding to the gap *C–E*; in the reverse movement it has been shortened by an equivalent overlap.

It should be realized that the movement of one wall or the other to produce a normal or a reverse fault is purely relative. Thus a normal fault may result from keeping the hanging wall stationary and moving the foot-wall block upward; or both blocks may move, in opposite

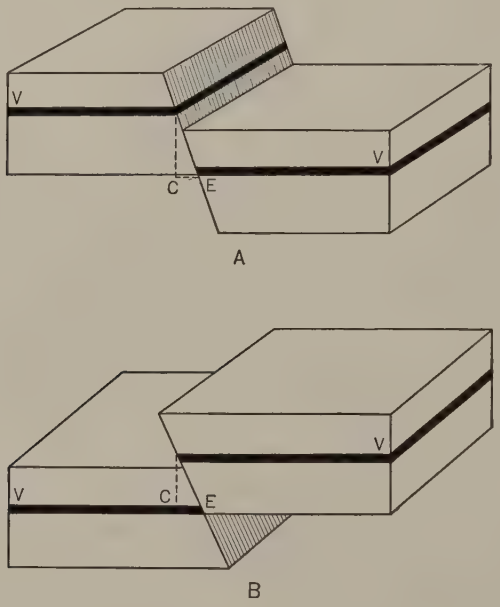

Fig. 226. — *A*. Simple normal fault. *B*. Simple reverse fault.

directions. In nature we see only the effect of faulting, and the actual character of the movement must be inferred. It is even possible that some faults described as normal or reverse were produced by horizontal slipping. The tilted layer *BB*, in Fig. 227, A, is offset by pushing the front block to the right. In a vertical section, such as might be exposed in the side of a canyon (Fig. 227, B), it appears that the hanging wall has moved down with relation to the foot wall; in other words, the fault would be described as normal. If the front block had been displaced to the left, a vertical section would show a reverse fault.

Components of Faulting. — It is often necessary for purposes of description or measurement to resolve a fault into component parts. On the diagram, Fig. 228, *AGEF* represents the horizontal plane and *A–F* is the strike of the fault. Let us suppose that the motion has been such

that a particle near A has been carried to the position B; then the line AB joining these two positions represents the *displacement*, or *slip*; no matter what actual path the particle may have followed, AB is the

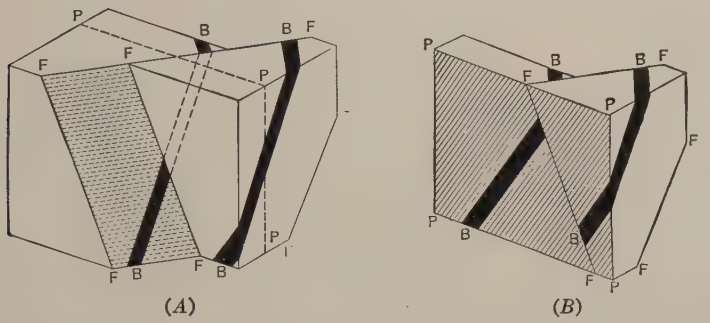

Fig. 227. — To illustrate normal faulting, as seen in a vertical plane PP, caused by simple horizontal shoving on the fault surface FF. The particular stratum B in PP (right-hand figure) appears to have slipped down. Modified from Ransome.

resultant, and its length the measure of the slip. The line AB, however, in order that it may be fixed and determined, must be referred to known axes or planes. The line AC gives the amount of motion in the horizontal direction at right angles to the strike FA, and this is known as

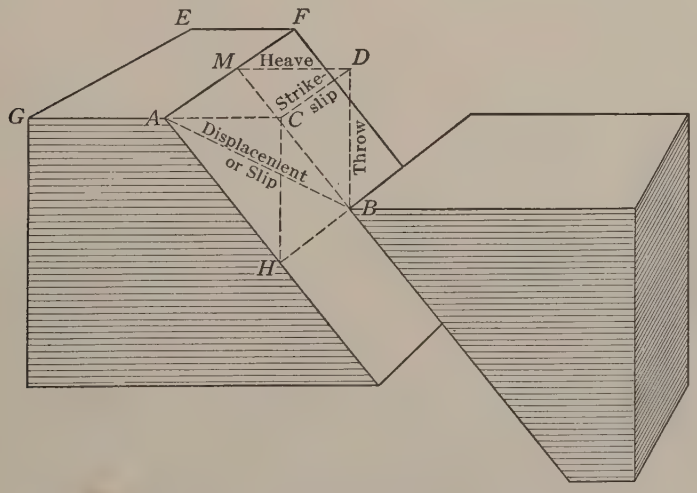

Fig. 228. — To illustrate and define the components of a fault.

the *heave* of the fault; the line CH is the amount of vertical motion and is called the *throw* of the fault; $HB(=AM$ or $CD)$, the amount of motion in the horizontal direction along the strike, is termed the *strike-slip* of

the fault. Thus there are the three right-angled axes *AC*, *DC*, and *HC*, meeting in the common point *C*, and these may be termed the *component axes* of faulting. The directions and intercepts on these axes being known, the displacement can be calculated, and the problem of the fault solved.

The heave and throw of faults are the components commonly recognized, because the dislocation is most easily seen in a vertical section at right angles to the strike of the fault, and in this section the particle *A* has apparently moved from *A* to *H*. The strike-slip is difficult to estimate in most faults and often it cannot be determined at all.

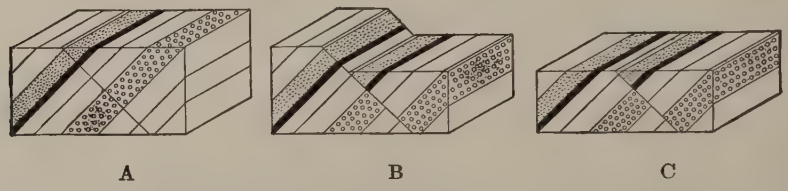

Fig. 229. — Illustrating strike faulting in stratified rocks. *A*, before faulting; *B*, after faulting, fault scarp still uneroded; *C*, surface levelled by erosion.

It is clear that a fault might take place without strike-slip, the movement being wholly down the dip of the fault surface. If the fault were exactly vertical, obviously there would be no heave; there might be strike-slip, but this also might be wanting and the fault would have throw only. Whatever the dip of the fault, conceivably the movement might consist of strike-slip only, without either throw or heave.

Faults in Stratified Rocks. — Although faults occur in all kinds and combinations of rocks, they show to best advantage in stratified beds,

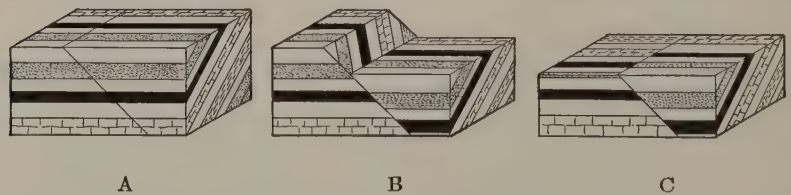

Fig. 230. — Illustrating dip faulting: *A*, before faulting; *B*, after faulting, fault scarp uneroded; *C*, surface levelled by erosion, showing offsets of strata.

on account of the strongly marked stratification which they disarrange. Certain terms are used to define faults in relation to the structure of the sedimentary beds. Thus in a *strike fault*, the strike of the fault and that of the strata are parallel, or nearly so, as illustrated in Fig. 229; *dip faults* cut directly across the strike of the strata, or nearly so, as shown in Fig. 230; *oblique faults* cut diagonally across the strike of the

strata. The figures show only normal faults; but they may also, of course, be reverse. The figures also indicate no real strike-slip, but in Fig. 230, C there is offsetting of beds, with a false suggestion of strike movement. Such abrupt offsetting of tilted strata is one of the surest

Fig. 231. — Repetition of formations by a normal strike fault. *a* and *a'* are parts of the same limestone member; *b* and *b'*, parts of the same shale member. The fault runs through *F–F'*, and dips to the right. Much of the hanging-wall block has been removed by erosion. Throw of the fault, 300 feet. Spotted Range, Nevada. (Longwell.)

indications of dip or oblique faulting. Strike faults are more difficult to perceive and may easily be overlooked; they may cause deception as to the thickness of strata by producing repetitions (Fig. 229, C and Fig. 231). Thus, in traversing strata the repetition of a certain set should lead to suspicion of strike faulting. On the other hand, strike faults may conceal strata after erosion has occurred. Thus in Fig. 232, which represents a reverse fault with later erosion, there is no outcrop of the stratum *A* at the surface.

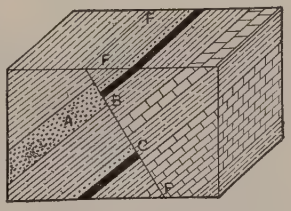

Fig. 232. — Illustrating concealment of strata by strike faulting and subsequent removal of the scarp by erosion.

The movement of one side of a fault on the other side may be attended by rotary or pivotal motion, as illustrated in Fig. 233. A fault of this nature is known as a *rotary* fault; or, if the displacement dies out gradually up to a definite point, it is a *hinge* fault. After erosion has levelled the surface, a fault of this kind is indicated by a pronounced difference in the strike and dip of strata on opposite sides of the break. Some faults pass gradually into monoclinal folds (Fig. 234).

The Magnitude of Faulting. — The scale on which faulting has taken place varies within the widest bounds. The displacement may be but a fraction of an inch, a number of feet, hundreds of feet, or even several miles. In the Plateau region of Arizona and Utah, several faults of

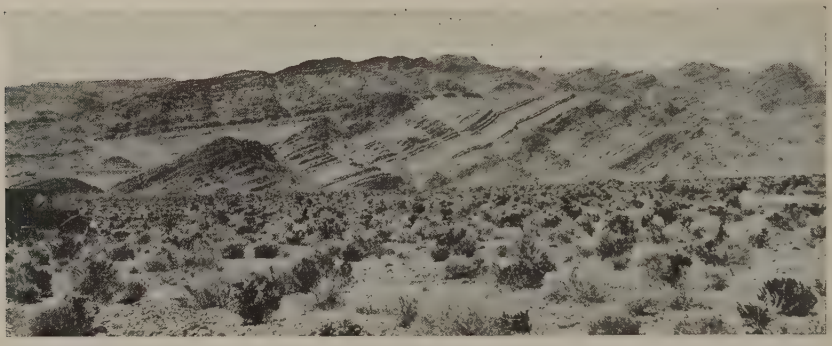

Fig. 233. — A pivotal or hinge fault. The fault runs from left to right, between tilted rocks and steep cliff on nearly horizontal strata. Displacement decreases toward the right, increases toward the left. Maximum throw about 3000 feet. Spotted Range, Nevada. (Longwell.)

great magnitude extend in a north–south direction, some of them crossing the Grand Canyon. Each of the largest fractures in this group can be followed 100 miles or more, and has a throw measured in thousands of feet. The Great Basin region presents the phenomenon of faulting on a colossal scale. In the area between the Sierra Nevada on the west and the Wasatch on the east, the crust is divided into huge blocks by gigantic fractures; and differential displacement of these blocks, together with erosion, has resulted in mountainous topography. A sunken tract of country due to downfaulting, or to uplift of adjacent areas, is called a *graben* (German for trough or ditch). Illustrations are the Jordan Valley and the Dead Sea, the great Rift Valley of Africa, and Death Valley in eastern California. An upstanding mass between two fault troughs or grabens is a *horst*.

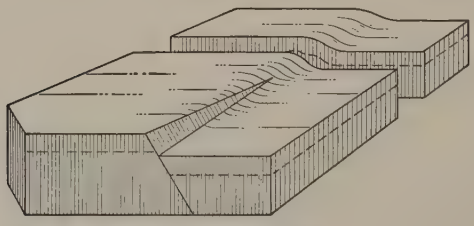

Fig. 234. — A normal fault passing into a monoclinal fold.

Thrust Faults. — Reverse faults are most common in those regions where crushing and folding of the Earth's crust have taken place; and the stronger the folding or crushing has been, the greater and more

evident the reverse faults are. They are especially evident in the stratified rocks of mountain regions, as in the southern Appalachians and in the Alps. Careful and detailed study of old eroded mountain areas has disclosed reverse faults of tremendous displacement, some of them with a comparatively low angle of dip, or even quite horizontal. A reverse fault that has a gently inclined fault surface is known as a *thrust fault*, or simply a *thrust*. Many of these are of such magnitude and importance that they are commonly considered by themselves as a special class of faults. The surface on which movement occurs is spoken of as the thrust surface, or less accurately as the thrust plane.

Such thrusts have been discovered and studied especially in the Alps, in northwestern Scotland, in the Scandinavian peninsula, in the southern Appalachians, in the Rocky Mountains from British Columbia to Utah, in southern Nevada, California, and in many other regions. The horizontal displacement of lower, older formations over younger rocks ranges from several miles up to 25, 30, or even 40 miles for individual thrusts. Figure 235 represents a portion of the great thrust along the front ranges

Fig. 235. — Section showing the thrust in northern Montana, whereby very old geologic formations of the pre-Cambrian are made to override the much younger beds of the Cretaceous. *BB* is the surface of thrusting; *D*, *D*, and Chief Mountain are erosional remnants of the pre-Cambrian resting on, and surrounded by, the younger Cretaceous. Displacement by thrusting observed, 7 miles; total amount unknown. (Generalized after Willis.)

of the Rocky Mountains in northern Montana. The deciphering of these great displacements is one of the triumphs of modern geological research.

Topographic Results of Faulting. — If a fault of considerable magnitude were to be formed suddenly, it would naturally be marked by displacement of the Earth's surface, giving rise to a cliff, or *scarp*. Numerous fault scarps are recognized, and some of them have been formed within historic time.

Such scarps in their original form may be called *initial fault scarps*. As the process of weathering and erosion works more actively in general on the uplifted side, the scarps become dissected and lowered, and slowly retreat from the fault line. Thus they pass through youthful, mature, and old stages. Finally, the difference in elevation on opposite sides of the fault line may disappear completely, and thus all topographic ex-

pression, initially due to faulting, may be obliterated. This would finish one cycle of erosion on a faulted surface (Fig. 236).

If now the whole region should be uplifted, without further displacement on the fault, and thus a new cycle of erosion initiated, then the

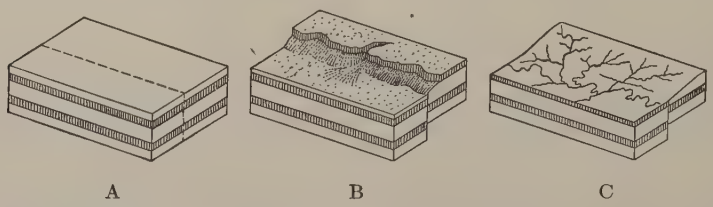

Fig. 236. — Shows the origin, development, and possible history of an initial fault scarp. *A*, block of strata containing two harder, more resistant intruded sheets of trap, before displacement. *B*, after faulting and some erosion; the fault scarp has become mature, and has retreated from the fault line. *C*, approaching the end of the first cycle of erosion; the fault scarp has been obliterated.

agents of erosion might find on opposite sides of the fault line rocks of quite different hardness and ability to withstand their attack. Therefore one side might be lowered so much more rapidly than the other as to leave the latter standing in a cliff or escarpment (Fig. 237). As such a cliff would result, not directly from the initial faulting movement, but from subsequent differential erosion, it deserves a distinguishing name, and has been termed by W. M. Davis a *fault-line scarp.* (See Fig. 237, which continues the history of the fault shown in Fig. 236.) The varying

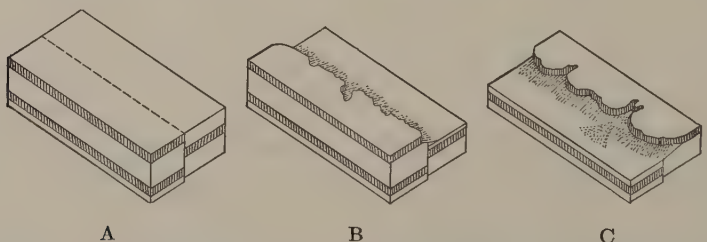

Fig. 237. — Possible development of fault-line scarps. *A*, the faulted block of the preceding figure commencing a second cycle of erosion; intruded trap sheets more resistant than the enclosing beds; uplifted block to the right. *B*, after erosion; a *fault-line* scarp has formed which faces *toward* the uplifted block. *C*, continued erosion has carried away the top trap of *B* and a new cliff has formed facing the other way, toward the sunken block; this fault-line scarp resembles the initial fault scarp (compare *B*, Fig. 236).

resistance to erosion on the opposite sides of the fault line determines naturally on which side the scarp will form, and a very long and old fault line might be marked in different parts of its course by cliffs facing in opposite directions. Finally, through the completion of another cycle of erosion these cliffs in turn might be worn away. The

influence of an old fault that reaches deep into the crust is evident in the topography so long as erosion of the area is active. Determination of the exact history of any fault is possible only after very careful geological study.

The east slope of the Sierra Nevada and the west slope of the Wasatch Mountains are fault scarps that have undergone a large amount of erosion. Nevertheless the faulting is recent in a geological sense. At the base of each of these ranges some of the movement has occurred so recently that initial scarps may be seen almost uneroded in the soft fans of alluvial material brought down by the streams. The last recorded movement near the Sierra Nevada took place in 1872. These faults have grown by successive movements, and the upper part of each scarp has suffered most from erosion.

The Plateau region, through which Colorado River cuts its way, is crossed by a series of great faults which are marked by prominent cliffs. These have been described as actual fault scarps; but most of them are fault-line scarps developed during a second cycle of erosion. Examples on a smaller scale are very common. Thus the sunken tract of sandstone and intercalated trap sheets between New Haven, Connecticut, and Springfield, Massachusetts, is divided into a series of tilted blocks by faulting. It has passed through at least one cycle of erosion, in which the initial fault scarps were eroded away; it is now in another cycle, initiated by broad uplift of the region without further movement on the faults.

Many old faults with vertical displacements amounting to thousands of feet are now practically unrecognizable in the surface forms. Obviously this relation in each case indicates erosion of great magnitude. Either there was a high fault scarp which slowly wasted away or the growth of the displacement has been so slow that erosion has kept up with it. This last suggestion is not altogether unreasonable, for we can scarcely imagine that the formation of great faults, with miles of displacement, has been a sudden process. Rather it results from gradual yielding of the shell of the Earth to forces brought to bear upon it during long periods of time.

The detection of faults that do not show any distinct topographic relief is possible through several kinds of evidence. The most common and obvious is the disturbance, or discontinuity, produced in the structure of the rocks, especially the stratified formations. In homogeneous masses of igneous rocks the recognition of faults is more difficult; yet even here discontinuity in certain features, such as dikes and veins, may lead to the discovery of faults and furnish a basis for measuring the displacement.

Origin of Faults. — The immediate cause of faults is comparatively simple and generally agreed upon; they result from strains set up in the outer shell of the Earth. Relief occurs by movement of rock masses, either along the surfaces of some previous fracture, or by the formation of a new one. Compressive strains give rise to reverse faults and thrusts; and it is natural that features of this kind occur commonly in areas of folded rocks, as folds also represent failure under compression. In regions of broad warping the rocks may be broken by torsion or twisting, which sets up tensional strains. After fracturing, displacement occurs by gravitative settling and readjustment of the fault blocks. Thus over wide regions where the strata are not otherwise disturbed, as in the Colorado Plateau, they are penetrated by fractures on which there have been great displacements. Also in the upper portions of uparching folds there may be tension and cracking, with subsequent gravitative settlement and faulting.

The *ultimate cause* of faulting evidently depends on those processes within the Earth which give rise to compressional or tensional forces and so set up strains in the lithosphere. These forces are most strikingly displayed in the formation of its chief features of relief, such as mountains and plateaus; and faulting may be considered only an attendant result of their operations. The forces themselves are hidden and can be inferred only from their effects. As the subject is obscure at best, and speculation must be guided by consideration of all available facts, it is best to postpone inquiry into the ultimate cause of crustal deformation until the structure and history of mountains have been discussed.

UNCONFORMITY AND ITS MEANING

Definition. — It is not uncommon to find, on examining the stratified rocks exposed in cliffs, valleys, and mountain sides, that one set of beds, whose composition, parallel position, and contained fossils prove them to be a continuously deposited series, rest upon another set of rocks, whose position and characters show equally well that they were formed at an earlier period and under other conditions. Thus in the diagram, Fig. 238, the layers of strata *d* have been deposited at one period and under one set of conditions; they are, therefore, spoken of as a *conformable* series of beds. Also beds of the series *c* are conformable among themselves; but it is quite evident that they are not conformable with *d*. Their attitude, and the abrupt termination of each bed upward, indicate that this series was tilted strongly and subjected to erosion before deposition of the overlying strata. The two series are *unconformable* with respect to one another, and the surface *ab* separating them is called an *unconformity*.

It should be understood clearly that the unconformable contact is a widespread *surface* and not merely the *line* exposed in a vertical section. Such a contact represents approximately an old land surface, or an old wave-scoured sea floor, and therefore is a clear record of erosion.

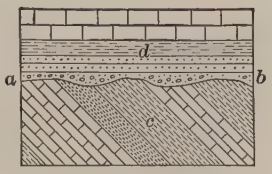

Fig. 238. — Section to show unconformity. The figure represents a vertical outcrop such as might be seen in the wall of a canyon. A conformable set of strata *d* rest unconformably upon another conformable set *c*; the line *ab* represents the unconformity.

It is not essential that the rocks beneath an unconformity be stratified. The lower formation might be composed of igneous rocks, such as granite; or of metamorphic rocks, such as schist or gneiss. If the surface *ab* can be identified positively as a record of erosion preceding the deposition of overlying strata, it is a surface of unconformity. If the contact between rock masses of unlike character is due to faulting or to igneous intrusion, the contact surface is not an unconformity.

Geologic History Revealed. — Suppose that *c* and *d*, Fig. 239, are two unconformable series of marine strata. Then the relationships represented in the diagram indicate a definite succession of geological events. First there was a long period of quiet deposition in which the beds of set *c* were laid down in a horizontal position, or nearly so, on a sea floor. The thickness of the beds, the kinds of rocks (limestones and shales), and the contained fossils constitute the record for this period. At some time later than the deposition of the youngest beds in this series, there was strong folding or faulting by which the strata were tilted steeply. Possibly the deformed mass reached to mountain heights. At any rate the deformation was succeeded by erosion, which planed the upended strata to a nearly even surface. From the section shown in the diagram we have no means of estimating the duration of uplift and erosion. Not only were there no records of the time formed in sediments in this area, but those of the previous period were wasted and obscured. Therefore there is a gap or "lost interval" in the geological record at this locality. Next in the geological history there followed a period of subsidence, when the eroded surface became sea bottom again, and received a new deposit of sediments, forming the conformable series of strata *d*. The events of this time are recorded continuously in the strata as before. Finally, after a second period of uplift, tilting, and erosion, the whole

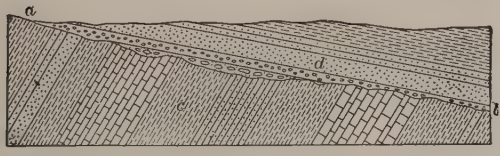

Fig. 239. — An angular unconformity with both series of beds tilted.

record is presented to us to be read so far as the evidence permits. The history here given may then be summarized as follows: first, deposition of strata; second, tilting, elevation, and erosion; third, subsidence and fresh deposition; fourth, final elevation, with tilting and erosion. If another subsidence should occur in the near future, the present land surface would mark a second abrupt break, above which a new set of horizontal strata would be deposited.

Relation to Kinds of Rocks. — Since an unconformity at the base of marine strata represents a submerged land surface, certain kinds of rocks are naturally associated with it. The sea advances inland, as a result of land submergence and of its own ceaseless gnawing at the shoreline. Where the land and sea meet there is generally a beach of the ordinary type, and as the land subsides this beach marches inland at the edge of the encroaching sea. Every part of the newly made sea bottom will have been passed over by this advancing beach; and all the superficial cover of the land — soil, pebbles, and rock decayed by weathering — is worked over by the advancing sea, and converted into beach material. The finer particles of the ground-up detritus are swept out to sea, and only the gravel and sand remain in the agitated waters near the shore. As an end result of this process, a continuous layer of conglomerate or coarse sandstone — the old beach material — commonly lies directly above the unconformity. The coarseness and thickness of this deposit varies according to the kind of rock composing the old land, the rate at which the sea has advanced, and other factors. A basal conglomerate or sandstone does not invariably accompany unconformities. More rarely the lowest deposit of the new marine series is shale or even limestone.

Classification of Unconformities. — Unconformities may be divided into two main groups. In the first, the lower formation, either by the tilting of the beds or by its composition of non-stratified rocks, shows at once its nonconformity with the series of beds above it. It is called an angular unconformity (Figs. 240, 241, 242, C and D) if the bedding planes of two stratified series meet at an angle. This term is good so far as it goes, but it does not cover the whole case, since the lower formation is not always composed of stratified rocks but may be of massive igneous or metamorphic rocks (Fig. 242, A and B). A more general term is needed, and an unconformity of this class is here termed a *nonconformity.*

On the other hand, the lower formation may be elevated, eroded, and submerged without material disturbance of the position of the beds. The old and the new formations will then have their stratification planes actually, or practically, parallel. This constitutes an unconformity of the second class, and as it is desirable that it should be distinguished from

Fig. 240. — Angular unconformity between tilted and eroded Paleozoic limestones and horizontal Tertiary conglomerate. Meadow Valley, Nevada, along the Union Pacific R. R. (Longwell.)

Fig. 241. — A closer view of the unconformity pictured above, in another part of Meadow Valley. (Longwell.)

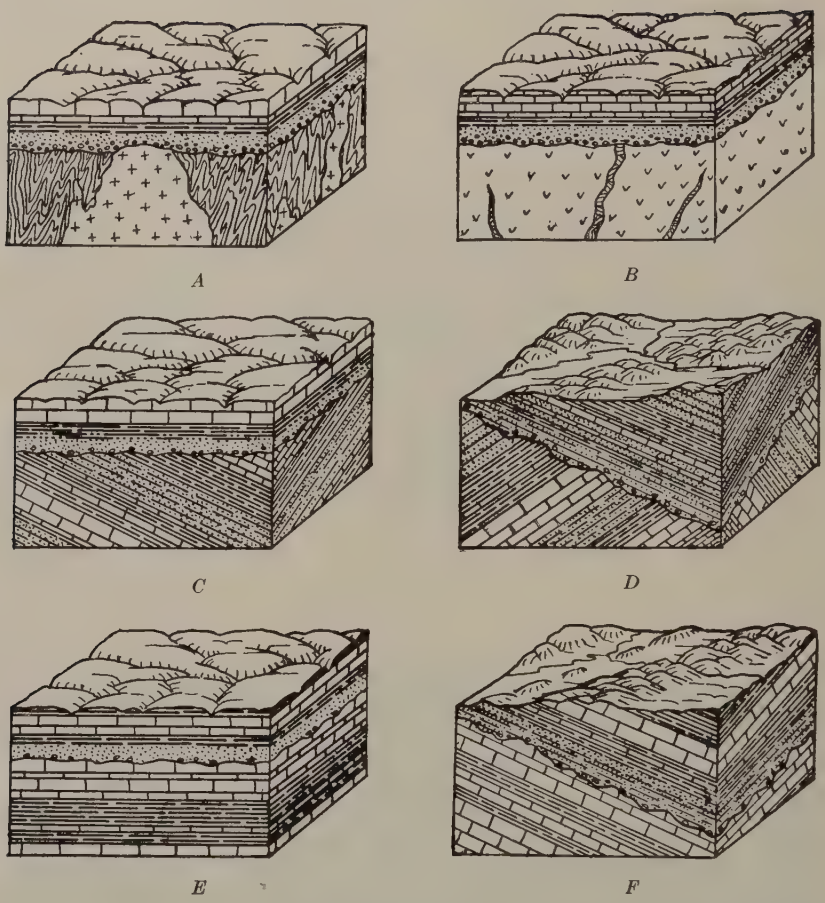

Fig. 242. — Diagrams to illustrate various kinds of unconformity. *A*, sedimentary strata deposited on metamorphic and igneous rocks; *B*, nonconformity on massive igneous rocks; *C*, angular unconformity between two sedimentary series, the later series undisturbed; *D*, angular unconformity, both series tilted; *E*, disconformity, both series of strata horizontal; *F*, disconformity, both series tilted together.

the other, it has been termed a *disconformity* (Fig. 242, E and F). We have then the following cases of unconformity:

Unconformity.

1. Nonconformity, two formations with visibly different structure.
   a. Lower formation of rocks nonstratified, or apparently so.
   b. Lower formation of stratified rocks, tilted with relation to overlying strata (angular unconformity).
2. Disconformity, two formations in parallel position separated by erosion surface.

Obviously the subject of unconformity is closely connected with the study of sedimentary rocks; but a full appreciation of its meaning requires some understanding of crustal disturbance. Strong angular unconformities record severe deformation, either by folding or by faulting; and widespread disconformities indicate warping movements that have involved large areas. A study of unconformities emphasizes the close relationship between crustal movements, erosion, and sedimentation. These ancient records, like the present wasting surface of the land, represent the continual struggle between deep-seated forces that produce irregularity, and surface processes that strive to keep the lands featureless and low.

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Gives excellent descriptions of structural features in general, with particular emphasis on types of structure that affect economic geological work.

CHAPTER XIII

EARTHQUAKES

It is difficult to think of earthquakes apart from their relation to human affairs. From the earliest recorded times the recurrent shaking of the solid ground, with consequent destruction on the surface, has been a cause of terror to man. Repeatedly, and in widely separated localities, populous communities have suffered great loss of life and property. Destructive earthquakes recorded during the brief span of human history are numbered by thousands. Geologic evidence indicates that violent shocks have been recurrent throughout the history of the Earth; and there is every reason to expect their frequent occurrence in the future.

The serious aspect of earthquakes from the human viewpoint is realized on review of some major catastrophes. September 1, 1923, approximately 100,000 lives were lost as a result of the Tokyo earthquake, and the estimated property loss exceeded $4,000,000,000. The shocks at Messina in 1908 and at Kansu, China, in 1920 were equally disastrous to life. According to report, more than a million and a half persons were killed by ten Chinese shocks between the eleventh and twentieth centuries; and Mallet, a profound student of *seismology* (from *seismos*, an earthquake) estimated that for the whole Earth at least 13,000,000 lives were lost through earthquakes in the course of 4000 years. Some activities of man himself, such as wars, or the operation of automobiles, result in a much higher death rate. Yet earthquakes are especially productive of fear, probably in part because they come without warning, and in part because their cause is more or less mysterious. Study of earth shocks from a geologic standpoint has dispelled a part of this mystery.

Cause of Earthquakes. — An earthquake is a trembling, or undulatory motion, in the more or less elastic rock shell of the Earth, communicated to it by an impulse or shock of some kind just as a bell is set in vibration by a smart tap on its side. The shock or impulse is evidently the immediate cause of the earthquake; but what is the origin of such shocks? Ancient philosophers who sought to explain natural phenomena by natural causes generally connected earthquakes in some way with the weather. Aristotle and Lucretius thought they were produced by winds rushing out of the Earth and leaving voids, with consequent collapse. Modern scientific evidence shows that shocks may arise from several

causes, most of which must be considered of minor importance compared with one major source, which appears to give rise to all great earthquakes.

One minor cause is in violent volcanic outbursts, like that of Krakatoa in 1883 and of Bandaisan in Japan in 1888; but earthquakes produced in this way are light in intensity and quite limited in extent. Moreover, many outbursts are not attended by any shocks, or at best by only feeble tremblings, such as occurred during the eruption of Mont Pelée in 1902. For a long time it was thought that volcanic action was an important source of earthquakes, and this idea is frequently revived; but the careful comparison of the two phenomena, especially in Japan, has shown that there is no necessary connection in occurrence between heavy earthquakes and volcanic eruptions. Instruments near Kilauea, in Hawaii, record numerous minor tremors — sometimes hundreds of them in a single month. Very few of these are accompanied by visible volcanic activity, although it is probable that shifting of magma at some depth is the principal cause of the local shocks.

Another minor cause of earthquakes may be the sudden caving in of subterranean cavities, or collapse of their roofs under the weight of superincumbent rock masses. This is most likely to happen in regions underlain by limestone, since large quantities of this rock are removed in solution by underground waters. It is possible, as has been suggested, that the earthquakes which in 1811 devastated the lower Mississippi valley, especially about New Madrid in southern Missouri, were partly due to this cause; though the area affected is so extensive and the effects of the earthquake shocks were felt to such great distances that caving probably was not the principal cause.

Fig. 243. — Map of a part of California, showing the position and extent of the fault line, A–A, movement along which produced the earthquake of April 18, 1906.

It has now been rather definitely settled that most of the major earthquakes result from the jar given by sudden yielding to strain in the Earth's crust. Such yielding may be by formation of a new fracture, or by

abrupt displacement along the walls of an already existent fault. In many areas visited by disastrous shocks the surface of the ground has been broken along fault lines and the amount of displacement is clearly indicated. Commonly these movements take place along old fault zones which bear the marks of repeated displacement. In California a great fracture zone can be followed almost continuously, by means of its

Fig. 244. — Trace of the fault concerned in the California earthquake of 1906. The deep soil above the bedrock broke irregularly. As movement was horizontal, no scarp was formed at this locality. (U. S. Geol. Surv.)

peculiar surface expression, from the southern part of the state northwestward for 600 miles. This feature, known as the San Andreas Rift, passes near the city of San Francisco (Fig. 243). On April 18, 1906, abrupt movement along at least 270 miles of this fracture caused a destructive earthquake. The length of this break is somewhat exceptional among historical earth movements; but similar breaks 25 to 50 miles long are not uncommon.

A careful study along the San Andreas Rift, after the rupture in 1906,

yielded valuable information on the nature and amount of the displacement. In general no scarp was made by the faulting, because the motion was almost entirely horizontal, parallel to the fault (Fig. 244). This fact was established beyond question by the offsetting of roads, fences, and other features that extended across the break. Some roads were cut cleanly across, and offset considerably more than their width (Fig. 245). The largest measured displacement was 21 feet. More commonly a part

Fig. 245. — Horizontal displacement of a road by movement on the San Andreas Rift in 1906. Before the earthquake the two offset portions of the road were in a straight line. The fault extends from left to right, directly across the road. (U. S. Geol. Surv.)

of the motion on a break of this kind is vertical and results in a steep scarp (Fig. 246).

If it is considered that the walls of a fault are pressed closely together, that movement is possible only by overcoming great frictional resistance, and that the displacement, once it occurs, takes place almost instantaneously, it is not surprising that powerful vibration is set up in the vicinity of the fault line. The exact nature of movement along the San Andreas Rift was made the subject of special study, and it is concluded that mass movement in the crust on opposite sides of the fault was in slow progress for years before 1906. Deformation in the rock was by bending, until the strain could be borne no longer and relief occurred by abrupt slipping along the old fracture. According to this idea, the sharp movement of

1906 was confined to a comparatively narrow belt closely adjacent to the fault, and did not involve the immediate shifting of great segments in the crust. The jar, resulting from the release of energy that had been accumulating for decades, was in effect a heavy blow which made the Earth tremble.

It is not to be supposed that a visible fault appears in every area visited by an earthquake. Commonly the direct evidence of crustal movement is wanting, especially in connection with mild or moderate shocks. It is

Fig. 246. — Displacement on a fault at Midori in the Neo Valley, Japan, in 1891. The former plain was broken, and one part dropped with relation to the other. Note vertical as well as horizontal offsetting of the road. (K. Ogawa.)

believed that actual displacement occurs very frequently at considerable depth and does not reach to the surface. This is a logical inference, as every break must be limited in extent, vertically as well as horizontally. Earthquakes of the first rank, however, are in a general way restricted to regions of active faulting, for which there is evidence at the surface. Some destructive shocks originate under the sea; but even in this event it may be evident from soundings that faulting has displaced the sea floor. For example after the Tokyo earthquake of 1923 it was found that an area in Sagami Bay had dropped more than 1000 feet below its former depth.

It was thought at one time that earthquakes were generated from a point at some depth below the surface, and this was called the *focal*

point, or *centrum*. The point immediately over this on the surface was called the *epicenter*. This latter point was determined by drawing concentric closed curves, called *coseismal* lines, on a map of the region through points of simultaneous arrival of the waves, as indicated by clocks (Fig. 247). By other mathematical methods the distance below the epicenter of the focal point was calculated. These methods led to discordant results for some earthquakes, and eventually to the discovery that for any one earthquake there might be several epicenters situated in a line, or that where earthquakes habitually occurred in a given region the different epicenters were situated along a line. Such a line probably represents a fault, even if there is no surface evidence of its existence. The terms centrum and epicenter still have value, although it should be understood that they are not points.

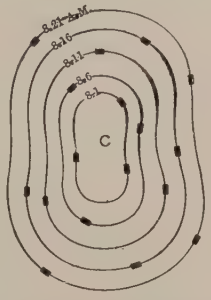

Fig. 247. — Map of coseismal lines. Black rectangles represent points at which the time of the earthquake was recorded accurately.

Effect of Shock. — It is important that we distinguish clearly between cause and effect in earthquake phenomena. The displacements shown in Figs. 245 and 246 are not the *results* of earthquakes as is commonly supposed; they represent the *causes*. The effect of the sudden movement along a fault is to set up vibrations that move outward from that place, and these constitute the earthquake, as it is perceived at a distance. Thus the earthquake is propagated as a series of waves in the highly elastic body of the Earth. When these elastic waves emerge at the surface the loose ground is thrown into rapid vibration, ordinarily with an amplitude not exceeding a few inches. In bedrock the amplitude is only a fraction of an inch. The actual amount of movement reaches a maximum in deep alluvium that is saturated with water. On terrane of this kind destructive effects are greatest. For example at San Francisco the devastation was most acute on the low alluvial flat near the bay. Buildings on solid rock, even much nearer the fault, suffered less damage. The contrast in behavior of the deep alluvium and of bedrock may be illustrated by striking sharply a bowl in which there is jelly. The bowl is set vibrating with resulting sound, but actual motion in the walls of the vessel is imperceptible. However the same impulse transmitted to

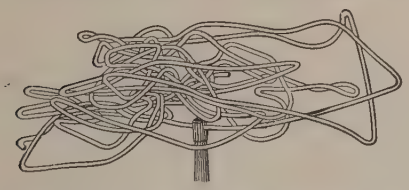

Fig. 248. — Wire model showing path traveled by a particle of matter during an earthquake; after Sekiya.

the jelly sets up longer waves that are visible. Excessive destruction on wet alluvium is caused by this larger wave motion as compared with the behavior of solid rock. In any kind of material the motion is not simple and rhythmic, but very complex (Fig. 248).

Recent Examples. — On August 31, 1886, the city of Charleston, South Carolina, was visited by a severe earthquake which did great damage. The shock was distinctly felt as far away as Chicago, a distance of 800 miles. This shock is of special interest, because from general considerations it does not appear that severe crustal disturbance should be expected at Charleston.

In 1899 a great earthquake took place in southern Alaska. As the region is mostly uninhabited the shock passed almost without notice at the time. Studies which have since been made show that considerable alterations in topography took place at the time of its occurrence, especially about Yakutat Bay. Marked changes were also induced in the great glaciers of this region by the shattering of the ice and by snow-slides from the mountains.

In August, 1906, the coast of Chile was visited by a severe earthquake, which did great damage in Valparaiso and other places. After-shocks continued for a long time while readjustment along the fault was going on. The west coast of South America is noted for its earthquakes, in connection with which notable elevation of the coast line has occurred.

One of the greatest disasters in modern times occurred on Dec. 28, 1908, when Messina and Reggio, cities on the narrow strait which separates Sicily from the mainland of Italy, were completely destroyed by a terrific shock. Evidently the area is in a zone of crustal weakness and readjustment, as severe earthquakes have occurred repeatedly.

The great Tokyo earthquake of 1923 is remarkable for the very large changes in topography that were produced by the crustal disturbance. Parts of the shore around Sagami Bay were lifted up as much as 6 feet. In addition to the large depressions on the floor of this bay, other parts of the floor were lifted several hundred feet, making shoals where there had been deep water.

These are only a few examples out of many that might be selected. Scarcely a day passes that shocks are not recorded from some part of the world by earthquake observatories.

Seismic Belts. — Although earthquakes occur in all parts of the world, they are most likely to happen in certain well-defined tracts, which lie in the two great seismic belts. One of these follows the western coast of North and South America, the Aleutian Islands, and the island groups along the eastern coast of Asia, and thus borders the Pacific Ocean on the east, north, and west. The other includes the Mediterranean, the Alps,

the Caucasus, the Himalayas, and continues into the East Indies, where it intersects the first belt at a large angle. (Figs. 249 and 250.) In a general way these zones coincide with the great volcanic belts (page 259); and this fact might appear to support the idea that volcanoes are an important cause of earthquakes. However, since the belts correspond

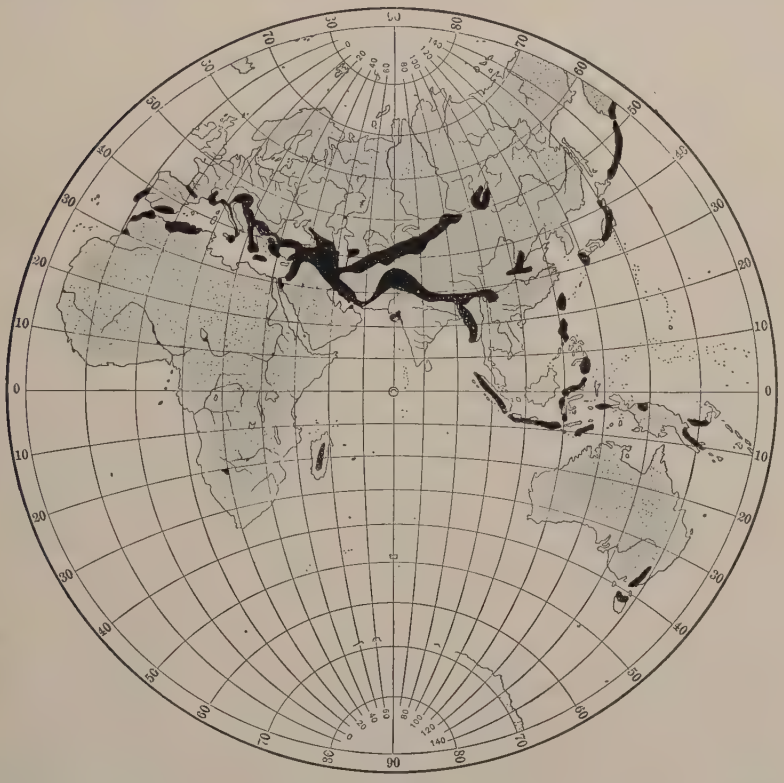

Fig. 249. — Map of seismic belts in the Eastern Hemisphere. On S. L. Penfield's stereographic projection. (Compare map of volcanic belts, Fig. 181.)

closely to young mountain systems and other marks of recent crustal movement, it is probable that both earthquakes and volcanoes have a common cause in this disturbance of the lithosphere. It is a notable fact that where the seismic belts lie directly along the continental borders, as on the coast of Chile and the eastern coast of Japan, the land descends sharply, without any broad intervening shelf, to great depths of the ocean. Some of these steep slopes descend into *foredeeps*, which are great troughs that appear to be sinking, while the bordering lands are rising. We conclude that these are zones of weakness in the Earth's

crust where strains are being constantly relieved by movements, and in which, therefore, earthquakes are continually recurring.

It is commonly thought that certain regions are practically exempt from danger of earthquakes because no real disaster has happened in them since they have been settled and cities have sprung up. It is true

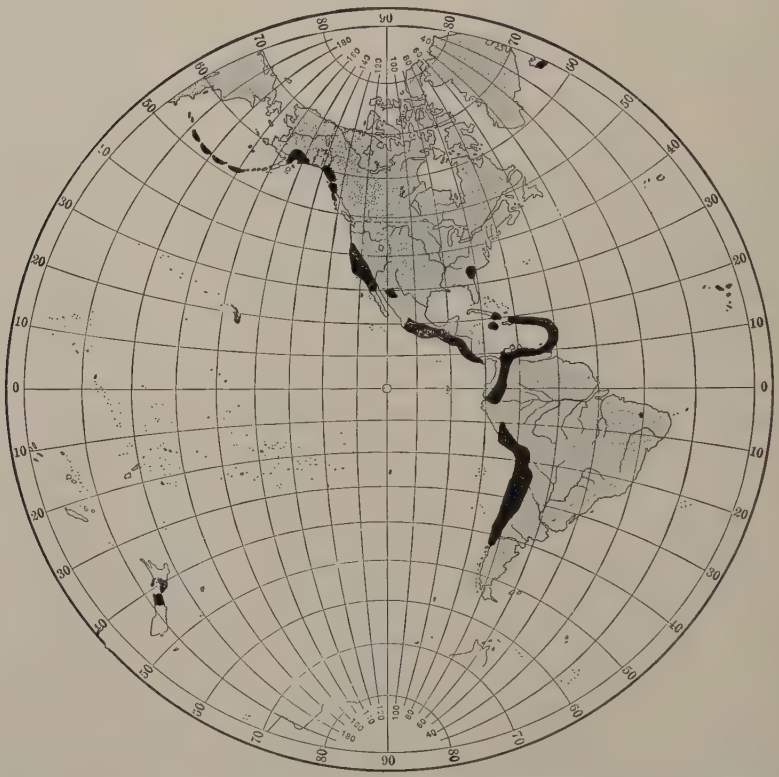

Fig. 250. — Map of seismic belts in the Western Hemisphere. On S. L. Penfield's stereographic projection. (Compare Fig. 182.)

that most of the Atlantic coasts, and large areas in continental interiors, are relatively free from earthquakes. The comparative stability around the Atlantic as compared with the Pacific is emphasized not only by the historical record, but by the existence of a wide continental shelf, which is in strong contrast with the Pacific foredeeps. However, the experiences of New Madrid in 1811 and of Charleston in 1886 are a warning that no locality may be entirely exempt. Even in New England, which is not recognized as a seismic tract, there has been an average of one perceptible tremor a year since the settlement of the country. Probably

none of the shocks has been of maximum intensity, although several have caused some destruction.

Submarine Earthquakes; Tsunamis. — The location of seismic belts suggests that many earthquakes originate under the ocean. Their occurrence beneath the sea is shown by shocks communicated to vessels on the surface above, and by rupturing of submarine cables. Since the invention of sensitive instruments by which it is now possible to record distant earthquakes and determine their location, it has been learned that a large proportion of all earthquakes occur on the floor of the Pacific. The most conspicuous mark of a submarine earthquake is the huge wave that commonly is generated in the ocean by disturbance of the floor. Such waves have long been known as tidal waves, a misleading name since they have no connection with the tide. They are now generally known to seismologists by their Japanese name *tsunamis;* or they

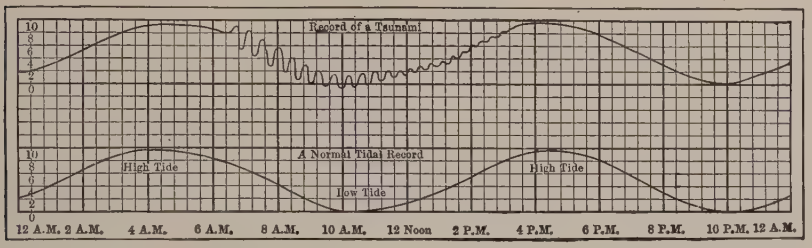
Record of a Tsunami
A Normal Tidal Record
High Tide
Low Tide
High Tide
12 A.M. 2 A.M. 4 A.M. 6 A.M. 8 A.M. 10 A.M. 12 Noon 2 P.M. 4 P.M. 6 P.M. 8 P.M. 10 P.M. 12 A.M.

Fig. 251. — Record of tsunami by tidal gauge. Vertical lines represent time spacing on the paper, driven horizontally by clockwork. Horizontal lines show height in feet as recorded by the rising and falling pencil of the gauge.

may be called seismic sea waves. Some are of immense size, measuring 100 or even 200 miles from crest to crest, and as much as 40 feet high. They are so broad that in the open sea they are not ordinarily perceived; but on approaching the coast they may pile up in huge breakers and, sweeping far inland, cause enormous damage and loss of life.

Lisbon in 1877, Japan in 1854 and in 1896, Peru in 1868, suffered from great and disastrous tsunamis. The number of victims of a single inundation of this kind has been as great as 20,000. These vast waves are felt over whole oceans and move with tremendous speed, — from 300 to 500 miles per hour. Those from Japan have crossed the Pacific in about 12 hours. At such distances their height may be only a few inches; but the ebb and flow of from 15 to 30 minutes, like small subordinate tides, are registered as wavy lines on the record of a tidal gauge (Fig. 251). These records make it possible to determine the size of the wave, since they give the period of oscillation, and since the velocity can be calculated from the time and location of the shock that caused the tsunami.

Recording Earthquakes. — Very delicate instruments have been invented, called *seismographs* (Fig. 252), which record the tremors due to distant earthquakes; and the study of these records has led to important geological conclusions. The principle upon which the commonest instruments are constructed is simple. If a heavy mass of metal be suspended like a pendulum, owing to its inertia it will remain for a time at rest when the shock arrives, while the bedrock vibrates beneath it. A pencil of some kind is secured to the suspended weight, and rests lightly on a paper or other medium suitably prepared to record the motions of the pencil. When the bedrock oscillates the heavy weight acts essentially as a steady point; but the vibration is transmitted to the pencil, and may be magnified as many times as desired by a simple mechanical device. If the paper, instead of being made fast, be a strip continuously carried along by clockwork, the pencil when at rest will draw a straight line upon it; when vibrations of the Earth occur the line will bend sinuously from one side to the other (Fig. 253). Such a record is known as a *seismogram*. Some instruments have, instead of a point or pencil, a small mirror that throws a beam of light upon a sheet of photographic paper. The seismogram is revealed only after the sensitized paper is developed.

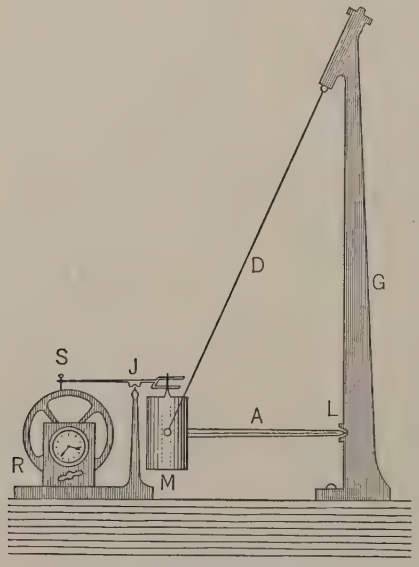

Fig. 252. — Diagrammatic representation of a seismograph. The upright post (G) is attached firmly to bedrock. The heavy weight (above M) is connected with the post only by a freely moving joint (L) and a flexible wire (D). Records are made by the stylus (S, J) on the revolving drum (R), which is driven by clockwork. (Sieberg.)

Although the principle of a modern seismograph is simple, in construction some of the instruments are rather complicated since they are arranged to record not only horizontal motion in two components, but also the vertical motion as well. It is from such records in three directions that the wire models like that shown in Fig. 248 are constructed. Since the intervals of time are marked on the moving paper, the instrument records the time of arrival of the shock and also the duration. The directions of diversion of the markers from their regular paths show also the direction from which the shock has come.

Seismograms. — The study of seismograms of distant earthquakes has led to the discovery that the main shock is preceded by smaller rapid vibrations which are recorded when the seat of disturbance is several hundred miles or more from the recording station. These are known as

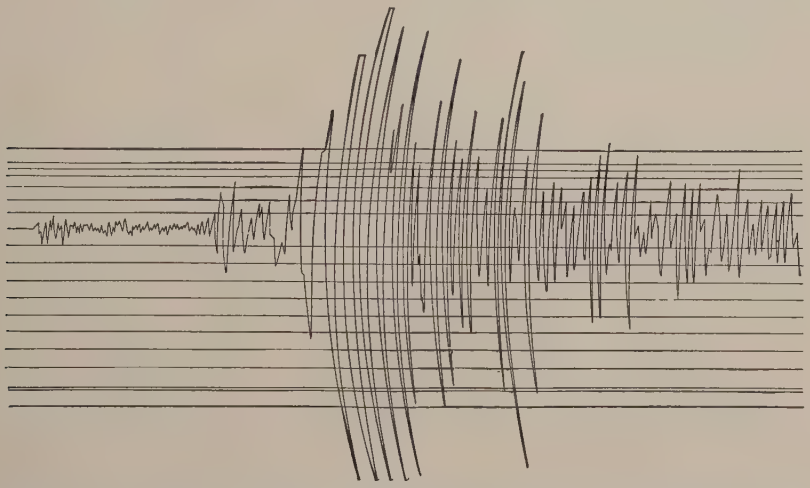

Fig. 253. — Record of the earthquake in Messina on Dec. 28, 1908, as shown by a seismograph in Göttingen, Germany, over 1000 miles distant. The actual vibration experienced by the instrument is greatly exaggerated in the seismogram.

the preliminary tremors. Thus a normal seismogram has the characters seen in Fig. 254. It has been determined that these preliminary tremors represent elastic impulses that come by the shortest path *through the Earth;* that is, in the general direction of a chord from the seat of disturbance to the recording station; whereas the later large vibrations represent those elastic waves that have traveled by a longer route over

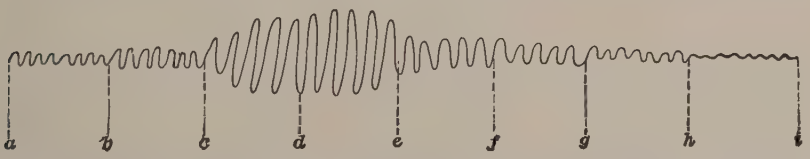

Fig. 254. — Seismogram of distant earthquake; *ab*, first preliminary tremors; *bc*, second preliminary tremors; *ce*, main shock; *fh*, later phases; *hi*, tail. (After Omori.)

the surface circumference. The first preliminary tremor (*a*, Fig. 254) is caused by a compressional or longitudinal wave (commonly known as the *primary*), which travels several miles per second; the other preliminary (*b*, Fig. 254) represents a transverse wave motion (the *secondary* wave), which travels at a distinctly slower rate. Therefore the time interval

between the two preliminary tremors is proportional to the distance traversed, and from this information the distance between the seat of the shock and the seismograph can be calculated accurately. By computing the distances from at least three separate stations, and drawing circles on a map with these distances as radii, the circles will intersect in a common point, which is the locus of the earthquake.

It is obvious that the circumferential or *long* waves will move out from the locus in opposite directions on any great circle of the Earth. If the locus of the shock should be exactly on the opposite side of the Earth from the seismographic station, these two sets of long waves would reach the instrument at the same time. Ordinarily one of the arcs is much longer than the other, and the second set of long waves arrives much later than the first; or if the distance is very great and the shock is slight, the second set may die out before reaching the instrument. Most seismograms record a succession of vibrations following the first long waves. Some of these later phases represent recurrent after-shocks, and others are due to complex reflected wave motions in the Earth.

Geological Deductions from Seismograms. — The fact that the preliminary tremors, which are supposed to travel through the Earth, arrive at distant points so long a time ahead of the main shock, cannot be explained alone by the shorter path traveled. The time interval shows that they are also propagated at a much greater rate of speed than the vibrations traveling in the outer shell of the Earth. The deduction from this is that they move in a more elastic medium than the superficial part of the crust. Moreover, the concordant results in different directions show that inside of the outermost layer, which we know is heterogeneous in composition, the Earth is homogeneous, or regularly arranged around its center in structure; or, if nonhomogeneous, the heterogeneous parts are relatively so small and numerous that different paths of considerable length through them give the effect of uniformity. Furthermore the average velocity increases with the distance of the recording station; thus the average rate of transmission along *sz*, Fig. 255, is greater than along *sy*, which in turn is greater than along *sx*. Velocities of the primary and secondary waves, as calculated from available earthquake data, vary with depth as shown in the following table:

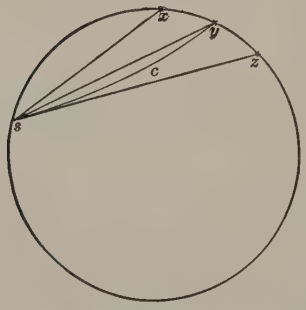

Fig. 255. — Paths of transmission of earthquake shock through the Earth. *s*, seat of shock; *x*, *y*, *z*, recording stations.

Depth Below Surface, in Miles	Velocity, in Miles per Second	
	Primary Wave	Secondary Wave
0	4.46	2.47
228	5.43	3.03
784	7.39	4.14
1680	7.95	4.25
2068	7.90	4.21

These results show not only that velocity increases down to a certain depth, but more and more slowly as the depth increases, and this would seem to indicate that the density and elasticity of the Earth increase with depth down to a certain region. It is a matter of considerable importance that the rate of propagation actually diminishes below a depth of nearly 2000 miles; and at still greater depth the secondary wave is not transmitted. Seismographs whose distance from the locus of an earthquake exceeds one-third the Earth's circumference (120°) receive no record of the transverse wave. The chord connecting the ends of an arc of one-third the circumference cuts the Earth's radius at its middle point. There is strong indication, therefore, that the Earth has an inner core, about 4000 miles in diameter, whose composition or state is different from that in the shallower zones. As will be explained in a later chapter, there is good reason for believing that the core consists chiefly of metal instead of rock. Some scientists have suggested also that this metallic core may be in the fluid state. This would explain the failure of the transverse waves to penetrate the core, as such waves are transmitted only through elastic solids.

From the fact that the rate of speed increases with the depth in the outer 2000 miles, it follows that the quickest path of wave transmission from the seat of shock to a distant station in this portion of the globe will not be a straight line, as from *s* to *y* in Fig. 255, along the chord of the arc, but will be a curved line slightly concave upward, somewhat like the line *scy*. In other words, by following this line the waves gain more in time in entering more elastic layers than they lose in distance, and hence seismologists generally assume that the path followed by the waves making the preliminary tremors at a distant recording station is curved. This is of some importance because, assuming the path to be straight and noting the fact that the preliminary tremors do not generally show in seismograms unless the distance is greater than 600 miles, the deduction has been drawn that there must be a rather sharp boundary between an outer rocky heterogeneous shell of the Earth and an inner

homogeneous core, and that, since the chord of an arc of 600 miles at its middle point is $12\frac{1}{2}$ miles below the surface, this must be the thickness of the outer layer. If a curved path is assumed, the thickness must be considerably greater. But Reid has suggested that the probable reason the preliminary tremors do not show in the records of "near" earthquakes is that instruments are not generally delicate enough to record and distinguish them from the principal shock, until distance produces appreciable time intervals. This view, expressed several years ago, is strengthened by recent investigations. Sensitive modern instruments situated less than 100 miles from epicenters have differentiated the preliminary tremors.

Geological Effects of Earthquakes. — There are several geological effects from earthquakes, but they are, comparatively speaking, of minor importance. The loose mantle of soil and other débris is often ruptured by the passage of the wave with the formation of fissures, which may be of some depth. A more important effect is the starting of landslides and avalanches in mountainous regions, through the jarring of the Earth. A variation in the flow of water from springs, or even the forming of new springs, has also been observed.

Much more important are the movements of the crustal blocks at the time of earthquakes; but as previously emphasized, these are the cause and not the effect of the shocks.

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CHAPTER XIV

METAMORPHISM AND METAMORPHIC ROCKS

Definition of Metamorphism. — In addition to the igneous and sedimentary rocks previously described there is a third class termed the metamorphic, whose distinctive characters are due to metamorphism. These rocks are the records of the remarkable transforming power of certain geologic forces that are at work within the Earth's crust.

Metamorphic means *changed in form*, and *metamorphism* is a general term for all those changes by which the original characters of rocks are more or less thoroughly altered, so that the component minerals or textures of the rocks are transformed into new minerals or textures, or both. Some metamorphic rocks pass gradually into those whose fossils and stratification prove them to be undeniably of sedimentary origin, whereas other metamorphic rocks grade into rocks whose characters show conclusively that they are of igneous origin. From these transitions we learn that some metamorphic rocks have been formed from sedimentary and some from igneous rocks. As we shall see later, some metamorphic rocks have also been derived from other metamorphic rocks.

The metamorphic changes may be so profound that the resultant product no longer resembles the rock from which it was derived but has become a new rock. Sedimentary rocks thus thoroughly *metamorphosed* are more coarsely crystalline, and the fossils that they may have contained and even the marks of stratification have been completely obliterated. Igneous rocks also, if severely metamorphosed, have lost their original distinctive features.

Limestone, for example, may be metamorphosed into coarse-grained marble with consequent loss of color and obliteration of the fossils that were in it; basalt may be converted into a green, slaty rock that gives no hint of its original igneous nature. Rocks that represent the stages of transition between the limestone and the marble or between the basalt and the green slate can be found; but under metamorphic rocks we include only those which have been so profoundly changed that their original outward characters have been either entirely obliterated, or nearly so, and distinctly new rocks have been formed. We say that it is the "outward" characters which are obliterated, for the bulk chemical composition of a rock as a rule remains unchanged during metamorphism.

The various changes that rocks undergo from the effects of *weathering* might in the strict etymologic sense of the term be classed as metamorphic. But they have been already discussed under the work of the atmosphere and the production of soils; therefore these agencies and the weathered rocks and soils that result from their action are not considered in this place.

KINDS OF METAMORPHISM

Three kinds of metamorphism are recognized: 1, contact metamorphism; 2, dynamic metamorphism; and 3, load metamorphism. Contact metamorphism is produced by the action of intrusive igneous masses on the rocks into which they were intruded. The effects produced are due chiefly to the heat supplied by the magma and the hot gases that issued from it when it consolidated. These effects are of course limited to the vicinity of the contact with the igneous mass, and hence the term contact metamorphism; but the term igneous metamorphism, by emphasizing the agency that produces the changes, is more appropriate. The most obvious form of contact metamorphism, probably the most impressive to the layman because coming more nearly within the ken of ordinary experience, is the conversion of a coal bed into a layer of coke by an injected sill.

Dynamic metamorphism results from the action of tangential pressure in the Earth's crust, as displayed in folding of the strata and in great overthrust faults. As metamorphism of this kind often accompanies profound dislocations of the crust, it is called by Heim, the great master of metamorphic geology, *dislocation-metamorphism.*

In places rocks appear to have been metamorphosed without either the intervention of igneous masses or of dynamic metamorphism. Here the cause of the metamorphism was apparently that the rocks were formerly deeply buried, having become depressed in the crust under a heavy load of overlying strata. The increase in temperature brought about by the deep subsidence has caused the development of new minerals and textures, and the heavy pressure due to the load favored the production of heavy minerals, such as garnet, which require the minimum volumes. As the Earth's own heat is the main factor here in bringing about the transformations, the process is sometimes called geothermal metamorphism.

The best established example of metamorphism of this kind is furnished by the German potassium-salt deposits. These salts were laid down in an evaporating arm of the Permian sea; they were many and of complex compositions; they were stable, however, under the conditions of moderate temperature then prevailing. Subsequently the basin in

which they were deposited slowly subsided and they became covered by 20,000 feet of sedimentary beds. The temperature of the deeply buried salt beds rose to that determined by their depth in the crust, and drastic rearrangements in the composition of the salts took place and many new minerals were produced.

Silicate rocks are far less sensitive than the potassium-salt minerals. Nevertheless, certain formations of pre-Cambrian age are believed to have acquired their metamorphic condition by load metamorphism. In places also the bottoms of synclines may have become so deeply downfolded that they have been subjected to load metamorphism.

Just as elsewhere in Geology, hard and fast lines do not separate the various kinds of metamorphism. Contact metamorphism may go on concomitantly with crustal folding; but we shall omit the description of these complex phenomena and devote our attention chiefly to the simpler forms of contact and dynamic metamorphism.

We shall begin with the contact-metamorphic rocks, for their origin is well established. As to the other metamorphic rocks, sometimes known as the *crystalline schists* in lieu of a better name, it is not always clear whether they are of dynamo-metamorphic origin or of load-metamorphic origin, or whether they were formed by the coöperation of contact metamorphism with dynamic or load metamorphism.

Contact-Metamorphic Rocks

As previously explained, the term contact metamorphism is used to denote the changes that are induced in rocks by the intrusion into them of a mass of magma.

The most noticeable effect of contact metamorphism is a baking, hardening, or toughening of the intruded rocks in a zone that surrounds the intrusive igneous mass. Since the intensity of these changes diminishes with distance from the intrusive mass, the contact-metamorphic zone is sometimes called somewhat figuratively a contact aureole.

Width of Contact Zone. — The width of the contact-metamorphic zone depends chiefly on the size of the igneous mass. The widest zones occur around stocks and batholiths. Around them the contact zone may be a mile wide, or even more; usually it is some hundreds of yards wide, but adjacent to a small intrusion such as a dike, it may be only a few feet. Lava flows produce at most a slight baking of the soils or rocks on which they rest.

The width of the contact zone around one and the same igneous mass may vary, for it is controlled by the configuration of the igneous body and by the attitude of the surrounding rocks. Thus in Fig. 256, section 1, as

a result of the lesser slope of the contact a wide zone is produced at *CD*, much wider than that adjacent to the vertical contact at *AB*. And in section 2 the beds at *F*, which slope toward the igneous rock, tend to have their bedding planes opened, and to furnish easy passageways to the emanations from the cooling magma. Since these emanations are the chief agents in carrying the heat and producing the metamorphism, it is clear that a broad zone *F* will be made on this side, compared with *E*, where conditions are reversed and a narrower zone must be formed.

Fig. 256. — Sections of intruded stocks and their contact zones. In 1, the breadth on the surface CD is greater than AB, depending on shape of igneous mass. In 2, the width F is greater than in E, depending on inclination of beds.

Contact Metamorphism of Rocks of Different Kinds. — The extent and the intensity of contact metamorphism depend very much, in addition to the factors already discussed, on the kinds of rocks surrounding the igneous mass. For our purpose here the sedimentary rocks may be divided into the three groups: *sandstones*, *shales* (and clays), and *limestones*. On pure sandstone the effect is rather small, though near the contact the sandstone may be changed into quartzite — a compact rock so firmly cemented that it fractures across the grains, instead of around them. Limestone is changed into marble, the masses of which may extend for considerable distances from their contact with the igneous rocks. Shales show the most notable and, generally, far-reaching results. The soft shales are greatly hardened, and near the contact they are converted into a rock known as *hornfels*, which to the unaided eye strongly resembles a black fine-grained igneous rock, such as basalt.

Coarse-grained igneous rocks, being the products of magmatic consolidation and therefore having already been at high temperatures, are generally but little affected by later intrusions. However, where volcanic rocks, such as basalts, have been invaded by granite batholiths, extensive and drastic metamorphism is produced.

The most interesting results are produced in limestones, especially impure, cherty varieties. Not only are the limestones turned into marble, but a great variety of minerals are newly formed in them, depending on the reactions that take place between the bases and acidic oxides present, especially lime and silica. Thus when the limestone is heated

above 500° C. the silica tends to drive out carbon dioxide,

$$CaCO_3 + SiO_2 = CaSiO_3 + CO_2$$

and calcite is changed into calcium silicate (wollastonite). If the limestone contains dolomite, then the following reaction may occur:

$$CaMg(CO_3)_2 + 2\,SiO_2 = CaMg(SiO_3)_2 + 2\,CO_2$$

and a pyroxene is formed, and carbon dioxide liberated. It is found that pyroxene is thus formed in the inner, hotter portion of the aureole, whereas an analogous compound, a white amphibole (tremolite), is formed in the outer, cooler portion of the aureole; thus these two minerals can be made to serve in a rough way as geologic thermometers.

If other impurities occur in the limestone, such as clay furnishing alumina, and iron oxides, many other new minerals will be formed. These more complex reactions may be illustrated by the following example:

Calcite	+ Clay	+ Quartz	= Garnet	+ Carbon dioxide	+ Water.
$3\,CaCO_3$	+ $H_4Al_2Si_2O_9$	+ SiO_2	= $Ca_3Al_2Si_3O_{12}$	+ $3\,CO_2$	+ $2\,H_2O$.

Thus a limestone that contained clay and sand as impurities may be changed into garnet with evolution of carbon dioxide and water.

Normal and Pneumatolytic Contact Metamorphism. — In the normal contact-metamorphic zone the effects produced are entirely owing to the heat given off by the igneous mass: the surrounding rocks become heated to a high temperature and new minerals are produced by the recombination of the elements already present in them, and consequently the chemical composition of the rocks remains unchanged. A marble, for example, has the same chemical composition as the limestone from which it was formed. If, however, the igneous mass during its consolidation gives off gases carrying iron, silicon, boron, etc., vastly different results are produced in the surrounding rocks. The most striking results are produced in limestones, because of the readiness with which limestones react with the magmatic gases. Many new minerals are formed, often beautifully crystallized. The resultant contact-metamorphic rock differs greatly in composition from the original rock: substances have been added in large amount by the gases that streamed from the magma. As gases are the means by which this transfer is effected and the reactions produced, contact metamorphism of this kind is termed pneumatolytic. If the gases contained iron, copper, or tungsten in notable quantity, valuable ore deposits may be formed in the limestones as incidental by-products of metamorphism of this kind.

"Crystalline Schists"

Most metamorphic rocks have *not* been formed by contact metamorphism, at least under the static conditions that prevailed while the rocks of the hornfels type were produced. So well known is this fact that the term metamorphic rocks when used without qualification refers to the great group that is not of contact-metamorphic origin. We lack a good distinctive designation for this group, and for want of a better name they are called the *crystalline schists,* in virtue of two characteristic features that most of them have: their obviously crystalline appearance and their foliated or schistose texture.

Many of these crystalline schists are the products of dynamic metamorphism, during which the rocks from which they were derived were forced by heavy differential pressure to flow in the solid state. As a result of this flowage there were developed either new textures, or new minerals, or both. Examples of such rock flowage are well shown in the Alps, perhaps the most marvelously folded tract in the world, where the limbs of the folds have become attenuated to one-tenth, even to one-hundredth of their original thickness and the arches of the folds have become enormously thickened: manifestly solid material has flowed from the flanks of the folds to the points of flexure.

Minerals of Metamorphic Rocks. — The minerals in rocks differ greatly in their ability to withstand the changes of temperature and pressure to which different metamorphic processes subject them. When new chemical and physical factors operate on them, they tend to change into new minerals, which are stable under the new conditions. The result of this adjustment is a metamorphic rock. The igneous rocks are characterized by a distinctive set of minerals, chiefly silicates, whereas carbonates and hydrated oxides as well as silicates are abundant in sedimentary rocks. Some minerals like quartz have a wide range of stability and occur in all three classes of rocks, but many minerals when subjected to metamorphic processes are converted into other minerals; thus carbonates are likely to be changed into silicates. Quartz, feldspars, mica, and hornblende occur in both igneous and metamorphic rocks, whereas garnet, staurolite, kyanite, talc, chlorite, and serpentine occur chiefly in metamorphic rocks.

Foliation. — ~~Most~~ metamorphic rocks are visibly crystalline to the unaided eye, and ~~they~~ which have a parallel arrangement of their constituent minerals. This parallel arrangement gives the rock a foliated texture (from *folium,* a leaf), and a rock having it is known as a foliate. By reason of it a rock tends to split, or cleave, more or less perfectly into flakes or slabs parallel to the foliation. A coarsely foliated rock is called

a gneiss, and one in which the foliation is well developed and closely spaced is termed a schist. Although foliation is the characteristic texture of metamorphic rocks, there are a few, such as serpentine, marble and quartzite, which commonly show no trace of this texture. In some rocks the parallel texture is straight or nearly so for considerable distances, as seen in Fig. 257; commonly, however, the banding is very much contorted, bent, or curled, showing the amount of deformation to which the original rocks were subjected (Fig. 260).

The foliated texture is due to the occurrence of the component minerals in separate layers or flat lenses, or to the parallel arrangement of prismatic or tabular minerals, such as hornblende or mica, or to a combination of both modes of arrangement. It is a result of the granulation and recrystallization to which the original rocks were subjected, and has been imposed on igneous and sedimentary rocks alike. The resemblance of the banding or lamination of schistose rocks to stratification led in the past to the erroneous view that all of them were derived from stratified rocks; that some have also been made from igneous rocks was learned much later.

Slaty Cleavage.— The cleavage exhibited by metamorphic rocks is most remarkably developed in slates, so characteristically that this variety of it is called *slaty cleavage*. Slates used for roofing, blackboards, and other purposes show this cleavage. The question of its origin has stimulated much investigation, along both experimental and mathematical as well as geological lines. From these studies it has become clear that the cleavage is the result of great pressure on the material, and that *the plane of cleavage is at right angles to the direction of pressure that produced it.* When fine-grained sediments, such as muds and clays, are subjected to intense pressure, the oblong particles in them tend to rotate so that their lengths are perpendicular to the direction of pressure; they also tend to become flattened perpendicularly to it. Important also is the fact that many of the platy or elongated minerals, such as the micas and the chlorites, have an excellent cleavage parallel to the flat directions. A considerable part of the minerals in the slate, such as the micas, were not originally present in the sediments, but were formed during metamorphism accompanying the pressure, and as they grew they set themselves in parallel orientation. All these features tend to give the rock a capacity to cleave readily in one direction.

The cleavage planes do not necessarily bear any fixed relation to the bedding planes. The beds were laid down in horizontal position and the direction of pressure is also horizontal or nearly so. Therefore the cleavage planes, being developed at right angles to this pressure, may cut the bedding at highly inclined or right angles. For the beds may become

Fig. 257. — Banded gneiss, Portland Township, Ottawa Co., Quebec.
(Geol. Surv. of Canada.)

folded before the pressure becomes intense, and the cleavage planes, being developed after the folding, will consequently intersect the bedding at various angles, though these cleavage planes themselves are all strictly parallel to one another (Figs. 258, 259, and 261).

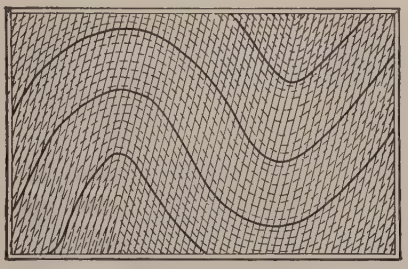

Fig. 258. — Slaty cleavage (represented by the nearly vertical lining) in folded beds.

Although most slates have been made from fine sediments, such as muds and clays, slaty rocks have also been produced by the shearing of fine-grained igneous rocks, such as felsites and basalts, and beds of volcanic ash. In the making of slates the original characteristic features of the rocks may become greatly distorted and even obliterated; thus fossils and pebbles in the stratified rocks and embedded crystals and other structures of the igneous rocks may be flattened into lenses or squeezed out into

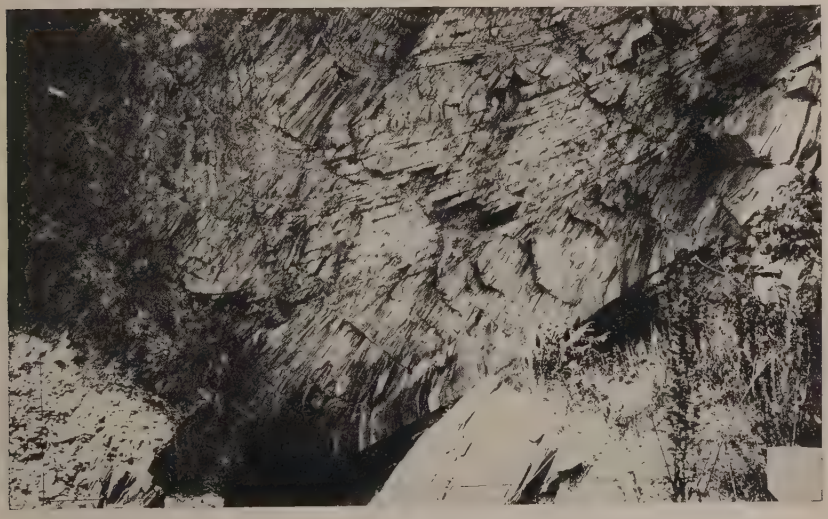

Fig. 259. — Slaty cleavage cutting at a high angle beds folded in a syncline. Slatington, Pa. (U. S. Geol. Surv.)

cylinders. Cleavage may in places be mistaken for original bedding, unless care is taken, and consequently the geological structure may be wrongly interpreted. It is sometimes important to indicate the cleavage on geologic maps, and this may be done by plotting its dip and strike, like that of a bedding plane. The important relation that cleav-

age bears to mountain ranges and their origin will be discussed under that subject.

Places of Occurrence. — Crystalline schists are widely distributed over the Earth's surface, and in some regions they are the only rocks exposed over extensive areas. Such extended tracts of metamorphic rocks were formerly said to be due to *regional metamorphism*, but inasmuch as that term explains nothing it is becoming obsolete. Its chief merit is its noncommittal character: it leaves open the question whether the metamorphism is due to igneous, dynamic, or geothermal activity, or to combinations of these activities.

Metamorphic rocks are particularly abundant in the most ancient (pre-Cambrian) formations the world over. In the United States they occur quite generally in New England, in the Adirondacks, and in a strip of country running from New York to Georgia. New York, Philadelphia, Baltimore, and Washington are built on metamorphic rocks. Metamorphic rocks are well shown in the inner gorge of the Grand Canyon of the Colorado, and in many other places.

The metamorphic rocks form also the cores of many mountain ranges, in which they have become exposed by denudation. The structure of these mountain ranges, as will be discussed later, includes folded strata, and the degree of metamorphism of the rocks is proportional to the closeness and intricacy of the folding. In other ranges, however, strata just as closely and intricately folded have not been metamorphosed. Hence mere intricacy of folding does not determine metamorphism. Probably the depth at which the folding takes place within the crust and the speed with which it occurs are the controlling factors. A relatively rapid rate of folding would generate the heat necessary to effect metamorphism.

The relation between folding, metamorphism, and mountain formation is so common that, where we find rocks intricately folded and very metamorphic, we assume that highlands once existed but have been eroded away, or, in general, that metamorphic rocks are exposed to our view only as a result of deep erosion. In conformity with this idea, metamorphic rocks are regarded as of *continental* origin, because they imply (when of sedimentary origin) the following sequence of events: deep erosion of a land mass to supply sediment; the deposition of this sediment in beds; the folding of the beds to effect metamorphism, perhaps with incidental production of a mountain range; and lastly renewed erosion to expose the metamorphic rocks. Such an array of processes can occur only on a great scale and therefore on and about continents; consequently when metamorphic rocks are found in place on Fiji, New Caledonia, South Georgia, and other islands, it is held that this proves that these islands are really remnants of continental masses.

Age of Gneisses and Schists. — Some of the facts previously mentioned led to the view that gneisses and schists must be geologically very ancient. This conclusion is now known to be but partly true. For although nearly unmodified beds of early geologic age that have been changed but little from their original horizontal position occur in Russia and in the upper Mississippi valley, on the other hand comparatively recent strata that have been greatly folded, such as those in the Alps and in some other mountains, have been strongly metamorphosed. The metamorphic condition of rocks depends wholly on whether or not they have been subjected to metamorphic processes; consequently the older the rocks are, the more likely it is that in the vast span of geologic time they have been affected by them: that is the large kernel of truth in the older view.

KINDS OF METAMORPHIC ROCKS

Introductory. — Because the metamorphic rocks are derived from igneous, sedimentary, and also from metamorphic rocks, it follows that there must be an extraordinary diversity of metamorphic products. Yet in the same way that we were able for ordinary purposes to gather the igneous and stratified rocks into a few groups, so we can consider the metamorphic under the few most important types.

Many sedimentary rocks are largely made of the detritus derived from disintegrated igneous rocks. In the process of disintegration it may happen that there is not much weathering and chemical change. Consequently the resultant sedimentary deposit will not differ much from the original igneous rock in chemical and mineral composition. Thus the red-brown sandstone (arkose) of the Connecticut valley, which contains much feldspar, has practically the same composition as the granite of the adjacent region. If such arkoses should become so thoroughly metamorphosed as to lose their original characters, they could not be distinguished from metamorphosed granites, nor could their former status be determined. From this example it will be clear that, while we can tell the origin of some metamorphic rocks at once, as is true of marble and quartzite, and can ascertain the origin of others after careful study in the field and laboratory, we are unable to ascertain the origin of many metamorphic rocks.

Classification. — It is possible to show in a general way the relation between the most common sedimentary rocks and their metamorphic derivatives in the following table:

Sediments	Sedimentary Rocks	Metamorphic Rocks
Gravel	Conglomerate	Gneiss, and various schists
Sand	Sandstone	Quartzite, and various schists
Silt and clay	Shale	Slate, phyllite, and various schists
Calcareous deposits	Limestone	Marble, and various schists

The igneous rocks, it will be recalled, are roughly divided into two main groups, the one chiefly composed of light-colored feldspathic minerals, and the other mostly of dark ferromagnesian minerals. We can illustrate in a rough way the relation between them and their metamorphic derivatives in the following table:

Igneous Rocks	Metamorphic Derivatives
Coarse-grained feldspathic types, such as *granite*, etc.	*Gneiss*
Fine-grained feldspathic types, such as *felsite*, *tuff*, etc.	*Slate* and *schists*
Ferromagnesian rocks, such as *gabbro* and *basalt*	*Hornblende schists*, various schists, and *serpentine*

Comparison of the tables will show that gneisses and schists may have diverse origins, as previously pointed out. Combining the results of these tables, we obtain the following main groups of metamorphic rocks, distinguished according to their mineral composition or by their texture, or by a combination of both.

GROUPING OF METAMORPHIC ROCKS

1. *Gneiss*, coarsely foliated rock.
2. *Quartzite.*
3. *Mica schist.*
4. *Slate* and *phyllite.*
5. *Hornblende schist; talc* and *chlorite schists.*
6. *Marble;* mixed carbonate-silicate rocks.
7. *Serpentine.*

Gneiss. — The most common variety of gneiss (pronounced *nice*) consists, like granite, of quartz, feldspar, and mica, but as the mica is arranged in more or less parallel planes, the gneiss has a rude cleavage. Some hornblende may accompany or replace the mica, and other minerals, such as garnet, may also occur, giving different varieties. Gneisses range in color from light to dark, and from fine to coarse in grain. Varieties transitional between granite and gneiss are very common. In

those gneisses made from conglomerates the original pebbles may still show as lenticular masses.

Gneiss is one of the most common of metamorphic rocks, and its many varieties have been formed under very diverse conditions: some under conditions of mild dynamic metamorphism in which the minerals have been merely mechanically deformed, and others under the conditions of most intense metamorphism, in which the mineral composition has been wholly reconstituted.

Some granite gneisses are called primary gneisses, because the foliated or gneissic texture has been assumed during magmatic flowage or by the

Fig. 260. — Contorted gneiss; Fullerton, Hudson Bay, Canada. (Geol. Surv. of Canada.)

shearing of magmas still in a partly consolidated or viscous condition. Batholiths that have been intruded during orogenic activity generally have foliated borders of primary gneiss. A view of beds of gneiss is seen in Fig. 260.

Quartzite. — Quartzite is a rock composed of quartz grains which are so firmly cemented that fracture takes place through the grains, instead of around them. Originally the rock was a sandstone and has been changed to a quartzite either (1) by filling the pore space of the original sandstone through the deposition of quartz from circulating ground water, or (2) by contact metamorphism, as already explained, or (3) by dynamic metamorphism. Quartzites formed as the result of the deposition of a quartz cement by ground water are not regarded as metamorphic rocks.

Quartzites of dynamo-metamorphic origin are commonly interbedded with gneisses and mica schists. The original pore space in the sandstone has been eliminated mainly by compaction and rearrangement of the quartz already present in the rock. Although quartzites of this origin are intimately associated with foliated rocks, they themselves rarely show foliation.

Quartzites are generally compact hard rocks of light colors — white, gray, reddish, or buff — and are likely to be of vitreous appearance.

Mica Schist. — Mica schist is the most widely distributed and important member of the great class of crystalline schists. The essential minerals are quartz and mica, and it is especially the mica that gives the rock its distinctive character. Different varieties of mica occur; the most common is a silvery white muscovite, which gives the rock a brilliantly spangled appearance, and the black mica — biotite — also is common. The micas are in irregular leaves or tablets with their cleavage planes oriented in the direction of the schistosity, and it is in fact this parallel arrangement of the micas that produces the extraordinary fissile character of mica schist. Some mica schists are dotted with dark-red garnets in well-formed crystals.

Mica schists are rocks that have attained a high-grade metamorphic condition. The initial material from which they were derived was an argillaceous sediment — a shale of some kind — and the transformation of the dull, amorphous substance of the shale to a brilliantly spangled mica schist is in a way as remarkable as the metamorphosis of a chrysalis into a beautiful butterfly.

Slates. — The origin of slates from fine-grained sediments, such as muds, clays, and ash deposits, by the action of compressive forces has already been discussed. While they may have various colors — red, green, gray, etc., — the most common one is dark gray to black, due to carbonaceous material in the original muds. As is well known, they are quarried for roofing slates, blackboards, and other purposes (Fig. 261). They are closely related to shales, but the distinction between them is that the cleavage or fracture of a shale is conchoidal, or "shelly" (whence its name, shale), and is parallel to the bedding planes; whereas in slates it is a secondary induced structure, which, as previously stated, is not necessarily parallel to the bedding.

Slates are metamorphic rocks of very low grade; some indeed — the clay slates — are very feebly metamorphic and are argillaceous sediments that have merely assumed a metamorphic texture, i.e., the slaty cleavage; others — the mica slates — have acquired new minerals as well as the slaty cleavage.

Phyllites. — Phyllites resemble slates, but their constituent mica is

coarser, which gives them a silky, glimmering luster. Most of them are transitional between slate and mica schist and are the results of an intermediate grade of metamorphism: more intense than necessary to produce a slate and not sufficiently intense to produce a schist. Some, like the ordinary slates, have been formed from sediments, but are more highly metamorphic. Others have been made from igneous material — felsite lavas, tuffs, etc., — by shearing and accompanying agencies of metamorphism.

Hornblende schist. — Hornblende schist is a common and typical variety of schist. It is generally dark green to black, and the parallel

Fig. 261. — Illustrates the occurrence of slates and cleavage. Slate quarries, Brownsville, Me. (U. S. Geol. Surv.)

prisms of hornblende, if not too large, usually give it a silky luster. Talc schist and chlorite schist are other common schists, in which talc and chlorite are respectively the predominant minerals.

Marble. — Marble is the metamorphic equivalent of the sedimentary carbonate rocks, such as limestone and chalk. Generally the marks of bedding and the fossils are effaced during metamorphism and the material is converted into grains of calcite that are visibly crystalline to the unaided eye. It is, therefore, harder, more compact, with purer colors, and takes a good polish. Just as there are ordinary limestones consisting only of calcium carbonate, and dolomitic limestones containing calcium-magnesium carbonate, $CaMg(CO_3)_2$, in variable quantity in addition to

the $CaCO_3$, so we have *calcite* marbles and *dolomite* marbles. Commercially this chemical difference is not important, but geologically it is of interest because the kinds of minerals that are likely to occur scattered more or less thickly through them are quite different in the two.

Marble is generally massive and shows no cleavage, even when it has been subjected to great stresses. As rocks go it is relatively plastic and flows under moderate pressures. If for example a dike or bed of schist is inclosed in marble that is forced to flow under differential pressure, the dike or bed, being brittle, will be ruptured and torn apart, and the marble will flow in between the dissevered fragments.

Pure marble is white, and the mottling, banding, and colors shown by ornamental varieties is due to impurities; the red and yellow tones to oxides of iron, the grays and blacks to varying proportions of organic matter. Besides being produced by dynamic metamorphism, marble is also formed by contact metamorphism.

Serpentine. — This name is given to a mineral, a hydrous silicate of magnesium, $H_4Mg_3Si_2O_9$, and also to a rock largely or entirely composed of it. The rock is usually greenish to black, soft, of a greasy feel, and massive, or without cleavage. Some of the blotched, lighter-green varieties are used as building and ornamental stones. Most serpentines appear to have been made by hydrothermal metamorphism (action of hot waters) on deeply buried masses of igneous rock rich in magnesia, such as peridotite, whereby the magnesium silicates change to this hydrated variety. Some impure dolomite marbles contain magnesium silicates (olivine, pyroxene, etc.) which may alter to serpentine. Some verde antique appears to be a serpentine-bearing marble of this nature.

RETROGRESSIVE METAMORPHISM

Metamorphic rocks, such as the series beginning with slate and comprising phyllite, mica schist, and garnet gneiss, represent stages in progressive or advancing metamorphism. Each successive rock in the series is the product of higher-grade metamorphism than the one that precedes it, chiefly as the result of adjustment to conditions of progressively higher temperature. A rock that has become adapted to the condition of highest metamorphic intensity may, however, be subsequently shifted into a new geologic environment, in which the conditions of stability are those of lower-grade metamorphic intensity. A garnet gneiss, for example, may be reduced to a phyllite along the base of a great overthrust block of the Earth's crust. Outwardly the resultant phyllite resembles a phyllite produced by progressive metamorphism, but internally, as shown by the microscope, it gives evidence of its former

high-rank metamorphic condition. By such retrogressive metamorphism many varieties of metamorphic rocks have been produced from other metamorphic rocks. By this process the already astonishing diversity of metamorphic rocks is greatly increased.

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CHAPTER XV

NATURE OF THE EARTH'S INTERIOR

Knowledge gained by study of the rocks at the Earth's surface impels man to speculate about the hidden interior. The materials and conditions that exist at great depth can never be known by direct observation. Openings in the solid Earth, such as mines and deep wells, are very superficial in comparison with the long radius of the globe. We infer that some of the rocks now exposed in the cores of old mountains were at a depth of several miles before erosion laid them bare; but even so they were always a part of the "outer shell," and we cannot be sure of their exact nature while they were under the relatively light load that has been removed in the course of long ages. If it were possible to make and maintain an opening to the center of the Earth, what would be revealed? Would the composition of material be found to change radically with depth, so that a large part of the interior would bear little or no resemblance to the superficial rocks? What would be the temperature at various levels? Would any large part of the interior be in a liquid condition? How would the materials at great depth behave under the enormous overburden? These are profound questions. They are not prompted by idle curiosity merely; they represent the goal of scientific investigations whose attainment would make clear many of the phenomena seen at the Earth's surface.

Whenever direct evidence is not available, science turns to the indirect and circumstantial. It is possible, with aid of this sort, to draw certain inferences that are sound; but if we seek to advance farther into the unknown, we must be content with hypothesis and speculation. Some suggestions can be accepted as probabilities, with the understanding that they may be found wanting after further investigation. It is well, at the outset, to state clearly the actual basis of fact, and to outline the methods of attacking the problem. In this way we shall avoid misconceptions, and be able to evaluate each suggestion on its merits.

INFORMATION AND METHODS

Size and Shape of the Earth. — The science of *geodesy*, which is concerned with exact measurement and mapping of the Earth's surface, has determined with precision the dimensions and the form of the globe.

This information, obtained by great labor through coöperation of scientific men in many countries, is of fundamental importance in problems relating to the Earth as a whole. It is known that the equatorial diameter exceeds the length of the polar axis by nearly 27 miles. As this figure is about $\frac{1}{297}$ of the diameter, it is stated that the *ellipticity* of the Earth is $\frac{1}{297}$. This departure of the globe from a true sphere is largely a response to rotation. Important deductions may be drawn from this known distortion of the Earth, considered with relation to other facts.

Gravity and Density. — By precise physical experiments we determine the "constant of gravitation." The method involves essentially the measuring of the force with which the Earth attracts a body of known mass. The result gives the total weight of the Earth; and since its size also is known, it is a simple matter to compute the average density. This value, arrived at by many experimenters, is 5.52; that is, an average sample of the Earth weighs five and a half times as much as an equal volume of water.

Direct determinations of density, using all kinds of rocks known at the surface, give an average value of 2.7. As this is less than half the density of the whole Earth, it appears that the interior must consist of much heavier material than the outer part. Other deductions will be discussed in a later paragraph.

Relation Between Density and Form. — In detail, the surface of the solid Earth is highly irregular. Continents stand well above ocean floors; plateaus, mountains, and deep troughs make irregularities of smaller order. On a small globe made to true scale these surface features appear insignificant; but in actual proportions some of them are large, and from a human viewpoint the irregularity is of the utmost importance, as without it the oceans would be world-wide.

Theoretically the surface of the Earth would be quite smooth if the material below the surface were uniform in its character. In reality, we know that the rocks exposed to observation differ considerably in composition and in density. Basalt and other dark-colored igneous rocks are appreciably heavier than granite. By geological investigation and by highly technical instrumental determinations, a strong probability has been established that the rocks underlying the oceans are in general denser than those composing the continents. Moreover it is concluded from careful geodetic study that the great mountain ranges of the Earth are composed of or underlain by material that is slightly deficient in density compared with the crust as a whole. It would appear from these facts that the larger surface irregularities are not haphazard, but have a fundamental cause; that differences in surface elevation are compensated by differences in rock density. Many important inferences and

some theories are based on this relationship. These matters will be discussed in a later part of the chapter.

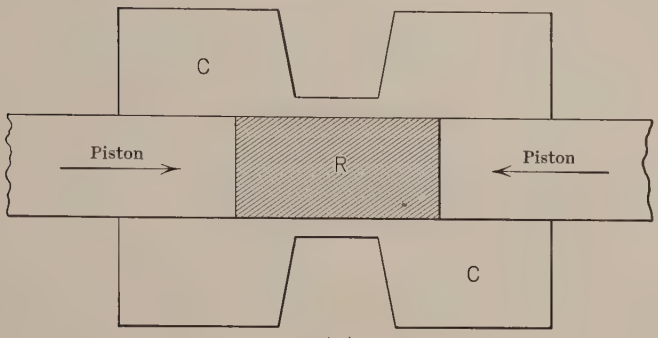
Behavior Toward the Moon and Sun. — The moon and sun exert a constant pull on the Earth, and the yielding of the oceanic waters to this pull gives rise to the tides. By careful and ingenious experiments it has been determined that the solid body of the Earth also responds to the tidal force. However, the yielding is very minute: about what would be expected if the Earth were composed of the strongest steel.

Another effect is produced by lunar and solar attraction on the equatorial bulge of the Earth, causing the globe to wabble slowly as it spins. Consequently the north pole of the heavens shifts slowly from year to year. This effect, called *precession*, is known precisely, and the forces involved can be calculated closely. The use of this information will be mentioned later.

Response of the Earth to Seismic Waves. — The transmission of earthquake vibrations through and around the Earth has been discussed. By study of the through-waves, the elastic properties at various depths can be deduced. In general, the Earth reacts to these impulses as a highly rigid, elastic body. The transverse or distortional wave in particular is of great significance, as this type of elastic wave is transmitted only through solid material. It is also significant that the velocity of both the transverse and the longitudinal waves decreases below a depth of about 2000 miles. This change appears to indicate either a different kind or a different state of material in the central core.

Other Facts About the Earth. — Volcanoes and hot springs indicate high temperatures at depth locally, and direct measurements in deep mines and wells suggest a universal increase in temperature downward. The rate of this increase varies between wide limits. In some places it is as high as 1° F. in 30 feet; in other places, as in the deep gold mines of the Transvaal, it is only 1° F. in 250 feet. The average for all observations is about 1° F. for 60 feet. Thus in wells that go to a depth of a mile and a half, the difference in temperature between top and bottom may be 125° F. or more. Some mines that reach down several thousand feet are uncomfortably warm.

The determination in recent years that all known rocks contain appreciable quantities of radioactive material is of great importance. It appears certain that the distribution of these materials must be confined to a comparatively shallow outer zone. Radioactive elements such as uranium and thorium break down at a constant rate, producing heat; and as rocks conduct heat away at an extremely slow rate, it can be calculated that the existence of these elements even at moderate depths would result in permanent fusion. Granites contain higher percentages of



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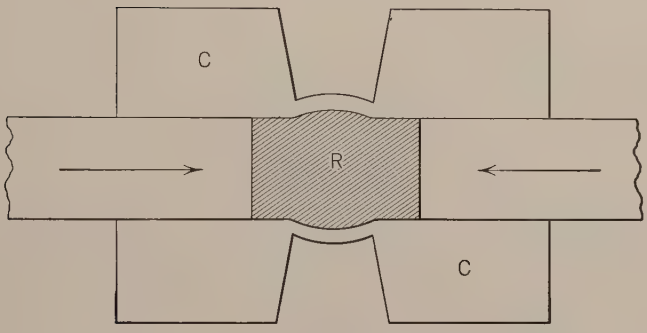

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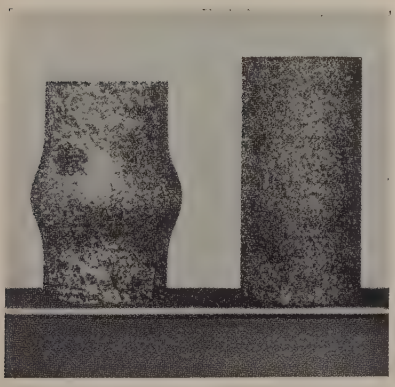

(*c*)

Fig. 262. — Rock flowage in laboratory test by Adams and Bancroft. (*a*) longitudinal section through steel cylinder (*CC*) with rock column (*R*) in place. Part of the steel wall was reduced to small thickness, as shown. (*b*) same after slow application of great pressure (up to more than 100,000 pounds per square inch) on the two pistons. (*c*) a rock column after and before the test.

radioactive substance than other known rocks, and hence it is argued that granites, which are the predominant rocks in the continents, are limited to shallow depth.

Behavior of Rocks Under Pressure. — Since ordinary rocks are brittle, they will fracture and crush under high pressures in the laboratory unless special precautions are taken. When a rock specimen is confined on all sides, and subjected to very intense compression, its size decreases slightly; in other words, rocks are somewhat compressible. If a marble core is fitted into a cylindrical opening in a strong steel jacket, and subjected to enormous pressure by means of steel pistons, in time the walls of the steel jacket are bulged outward. By cutting away the jacket it is found that the marble core has been shortened and thickened, *but not crushed* (Fig. 262). Evidently it has *flowed* slowly, as if it were a plastic substance. Since rocks deep in the Earth are confined under high pressure, it is argued that under certain conditions these rocks are deformed by plastic flow. Strength and rigidity are only relative terms.

INFERENCE AND HYPOTHESIS

Density Distribution. — It is certain that the outer part of the Earth consists of much lighter material than the average. Several assumptions might be made, however, as to the arrangement of light and heavy substances between the surface and the center. For example, it is possible that light rocks, like granite, form a shell a few miles or tens of miles deep, and below this shell the Earth is composed of very heavy rock with a uniform density of approximately 6. Again, it may be assumed that from the outer zone of low density there is a gradual and progressive increase to a density of 9 or 10 at the center. By either of these arrangements the average density of 5.52 might result. Besides these two suggestions a number of other assumptions as to distribution of density might be made, all of them consistent with the known average density.

Fortunately, there are other checks to guide us in attacking the problem. Any assumed distribution of the density must harmonize with the mathematical and mechanical knowledge to which reference has been made above. The small degree of flattening at the poles of the Earth suggests that a large percentage of the mass is concentrated in the central portion; and the same suggestion is evident from study of the precessional effect. These considerations, therefore, favor the assumption that density increases slowly downward in the Earth, and that the central core is composed of very heavy material.

Once this point is reached in the inquiry, another problem presents itself. Is the increase of density toward the center of the Earth due

wholly to compression under enormous weight, or is there a concentration toward the center of metals that normally are heavy? Some students of the problem have argued that compression of ordinary rock material is a sufficient explanation. At a depth of one mile, each square foot of rock bears a weight of 450 tons; and with each additional mile the pressure is increased by more than this amount, as the material grows progressively denser. Near the center, pressures amount to more than 3,000,000 tons per square foot. Without question such intense compression has an effect in compacting the material. Since no pressures that are possible in laboratory experiments can even remotely simulate conditions deep within the Earth, we cannot state positively how much compacting can result. However, from various lines of evidence and reasoning it appears unlikely that ordinary rocks can be compressed sufficiently to give the high average density of the Earth. Therefore many scientists favor the view that the core of the Earth is in large part metallic.

Which of the metals is most likely to exist in such abundance in the Earth? The most satisfactory answer to this question comes from consideration of the material that reaches us from outer space. The majority of known meteorites are composed of nickel and iron, and the others consist of dark-colored, heavy rock. Since these bodies probably were derived originally from the sun, along with the material in the planets, it is argued that they suggest the composition of the Earth. Adopting this argument, and using all available information, certain scientists have postulated the following arrangement in the Earth:

(1) An outer layer about 35 miles thick, in which the material changes gradually from granite to dark rock somewhat heavier than gabbro; (2) a zone extending to a depth of nearly 1000 miles, consisting of peridotite, with a density ranging up to more than 4; (3) a zone reaching to a depth of nearly 2000 miles below the surface, in which peridotite is gradually replaced by iron or nickel-iron, with a density of 9.5 at the base of the zone; and (4) a central core of nickel-iron, with density about 10. A part of the increasing density with depth is attributed to compression under load, but a larger part to materials that normally are heavy (Fig. 263).

It should be kept in mind that this suggestion is merely a hypothesis, no part of which is subject to proof at present. It is only an attempt to give a picture that is consistent with all known facts.

Temperatures at Depth. — The average change in temperature for a given unit distance is known as the *temperature gradient.* If the gradient determined in mines and bore holes should continue downward unchanged, the temperature of the center would exceed 350,000° F. Aside from the fact that this figure seems inconceivably high, several consider-

ations make it improbable that temperatures in the interior are so excessive. The length of the deepest opening used in estimating the gradient is less than $\frac{1}{2500}$ the Earth's radius; and the value obtained for this thin skin cannot be accepted with any confidence for the whole body. Rocks such as granite are extremely poor conductors of heat. Therefore if the temperature at a depth of several miles should be high, say 1000° or 1500°, heat would flow out very slowly, and the change in temperature for each 100 feet would be considerable. It is altogether likely that the heat conductivity improves with depth, both because high pressure compacts the rocks and because metallic substances, which are notably

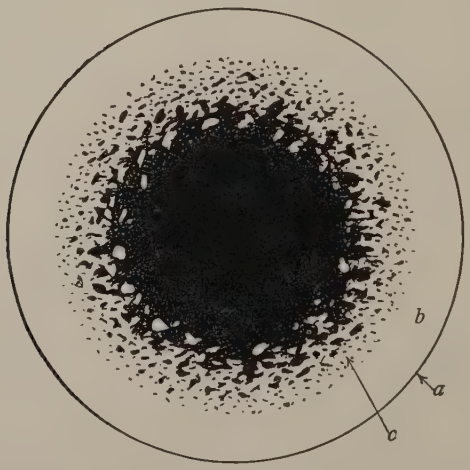

Fig. 263. — To suggest the concentration of metallic substances toward the center of the Earth. (*a*), outer zone of granite and basalt; (*b*), zone of peridotite; (*c*) peridotite mixed with nickel-iron. The black central portion represents a core of nickel-iron. (Adams and Williamson.)

good conductors, probably grow more important in the deeper zones. Thus, whatever the temperature of the deep interior, it is likely to be distributed more uniformly than in the outer zone. To express this conception another way, let us imagine that the core of the Earth is very hot. If the structure and composition are as represented in Fig. 263, the heat will flow out rapidly through the zones made of metal and of heavy rock, producing a nearly even temperature; but on reaching the outer zones, made of granite and other poor conductors, the flow of heat will be checked, and the temperature will fall rapidly in approaching the surface. Therefore the gradient in the shallow zone probably is larger than for any other part of the globe, and cannot be used to calculate the temperatures at great depth.

It is even probable that much of the heat conducted to the Earth's surface and lost by radiation does not come from great depth, but is generated in the shallow zone. The presence in granite and other known rocks of uranium and other radioactive elements has been mentioned. These elements disintegrate slowly but continuously, with evolution of heat. The possible significance of this process in connection with igneous activity has been discussed previously.

Whatever may be the true value for the average temperature gradient, many lines of evidence suggest that temperatures in the interior are high. Probably they are above the melting points for the materials under surface conditions; but as earthquake waves testify that no considerable part of the globe is molten in the outer 2000 miles of its radius, it is certain that heat is not great enough in this portion to exceed the "critical temperature" under the great pressures that prevail. Locally this control by pressure may be overcome in one of two ways: by actual decrease of the pressure, or by unusual concentration of heat. According to our conceptions, pressure may be relieved only at shallow or moderate depth, by local arching up of the superficial rocks or by deep fracturing. The most conceivable reason for exceptionally high temperature in the rocks is the presence of radioactive substances in unusual amount. As our reasoning confines these substances to the outer part of the Earth, it appears that conditions for actual liquefaction of rocks are favorable only at shallow or moderate depths. From local pockets of magma generated by any cause the surplus heat is dissipated in time, either by volcanic activity or through movement of the magma into higher and colder rocks. Thus the temperature finally is reduced below the critical point, and the solid condition is resumed.

The condition of matter in the central core is still a subject for varied speculation. From the testimony of earthquake waves we reason that the core is not an elastic solid. Some scientists argue that the interior heat is sufficient to keep the metallic core fused in spite of the enormous pressures at that depth.

Isostasy, or Equilibrium in the Crust. — The Earth does not have the form of a cube or a pyramid or other angular figure, for a good mechanical reason. Rock is not indefinitely strong. Under great strain it adjusts itself to a condition of equilibrium, as does any other material. If the Earth could by any means be forced momentarily into a greatly distorted shape, the laws of Nature would bring about restoration to a figure of equilibrium. This figure would be essentially a sphere if there were no rapid rotation; but the spinning on an axis makes the figure an oblate spheroid, with the degree of flattening determined by the rate of spinning.

How high may a mountain mass rise above the general surface, or how

far may an ocean "deep" lie below it? Certainly not to unlimited height or depth, for then the natural law would be violated. The strength of rock is sufficient to bear considerable strain, but there is a definite limit, as may be demonstrated by laboratory experiment. Is this limit ever exceeded in the outer part of the Earth? In geology we see the evidence that great masses of material are shifted from one part of the surface to another. Rocks are folded and crowded together by horizontal movements in mountain zones; continents and mountains are eroded, and the débris is piled along the continental margins; more than once vast quantities of water have been removed from the oceans and heaped upon the continents in the form of thick ice sheets. These transfers of material tend to change the form of the spheroid, and undoubtedly set up great strains. Is there some mechanism for adjusting these strains?

This view of the subject is altogether deductive. There is also a more practical avenue of approach, through geodetic measurements and geo-

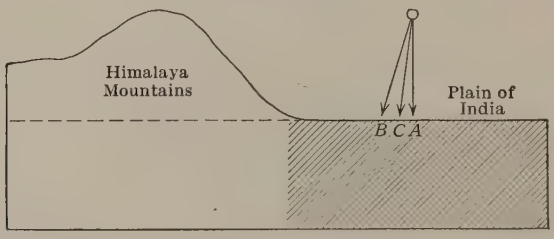

Fig. 264. — Effect of deficient density in a mountain mass. *OA* represents the true vertical; *OC* the actual position taken by a plumb line; *OB* the calculated position of the plumb line assuming rock of uniform density under mountains and plain. Angles and vertical scale much exaggerated.

logic evidence. The geodesists use two important methods of investigation. A plumb line suspended on a plain near a great mountain front is attracted laterally by the upstanding mass, and therefore does not point exactly toward the center of the earth (Fig. 264). Knowing the volume of the mountain unit, and assuming that the rocks in and beneath the mountains have the average density of normal rocks at the surface, we can compute accurately the amount of lateral attraction to be expected. In actual experiments the deviation of the plumb line from the vertical is only a small fraction of the expected value; and therefore we reason that the rocks in or under the range are abnormally light. This conclusion is checked by precise measurements of gravity on the high and low areas, by the use of a delicate pendulum. The rate at which the pendulum vibrates is governed closely by the intensity of gravity, which varies at different localities with latitude, height above sea level, and other known factors. By considering these factors,

geodesists can calculate closely what the value of gravity at any station should be; but for mountain stations it is found in general that the calculated values are too high. Again the result is explained by assuming that the excess *volume* of material represented by the mountains is offset by deficient *density* beneath. It is inferred, therefore, that high areas such as continents or mountain chains are more or less balanced against low areas such as ocean basins or low plains. The term *isostasy* (from the Greek meaning " equal standing ") is used for this supposed condition.

A simple illustration of the principles involved in isostasy is given in Fig. 265. The different metals vary considerably in density. Therefore if blocks are taken with the same weight and cross section, blocks of the light metals are considerably longer than those of heavy metals. All of these float in mercury, which is 13.6 times as dense as water. As the blocks have equal weight, they sink to the same depth, leaving the top

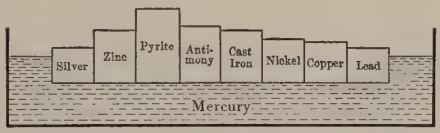

Fig. 265. — Diagram to illustrate an irregular upper surface on floating blocks of differing density. The blocks are equal in cross section and in weight, and therefore sink to equal depth. This is an ideal illustration of one isostatic theory, which assumes that differences in altitude between major features of the Earth's surface are compensated by differences in density down to a certain level, below which the density is essentially uniform. (Bowie.)

surfaces at irregular heights. The longer blocks might be taken to represent mountains and plateaus; the shorter, low plains or basins. It is not to be understood, of course, that the Earth's crust is divided into definite blocks, or that there is a liquid substratum at some depth. The illustration is highly artificial, and any attempt to press the comparison closely will result in misconception. It is intended merely to emphasize the principle of mass balance.

Figure 265 may suggest that great differences in topography reflect large variations in kinds of rock at the surface. From geologic evidence, however, it appears that the commonest rocks everywhere in the continents are granites. Therefore it is possible that mountains represent merely a local thickening of the granitic crust. This conception is illustrated in Fig. 266, in which all the blocks are of copper, but of different lengths. Irregularities at the surface are reflected by similar inequalities reaching downward. This general arrangement applied to the Earth would satisfy the principles of isostasy as well as the conception in Fig. 265, and would accord better with geologic evidence.

Suppose now that a portion is cut from the top of one block (in either figure) and placed on another. Equilibrium is disturbed, and adjustment takes place by sinking of the loaded unit and rising of the other. In a liquid this adjustment is immediate and perfect. How can the principle apply in the Earth? We are satisfied that no continuous zone of liquid rock exists; but it is known that solid rock behaves as a plastic substance under high pressure. The laboratory proof has been mentioned, and circumstantial evidence of rock flowage is seen in metamorphic rocks exposed by deep erosion. It is inferred, then, that any overloaded part of the crust sinks, displacing the deep rocks by plastic flow and thus causing lighter parts of the crust to rise. The operation is

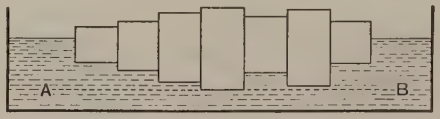

Fig. 266. — Copper blocks, equal in cross section but unequal in length, float in mercury. They sink to unequal depth, and also rise to unequal height. It can be shown that if the mercury directly under each block down to the line *A–B* is included, all the blocks are of equal weight. According to this conception of isostasy, a continent consists of a granite shell, essentially uniform in density but with variations in thickness. The thinner portions form low plains, whereas thicker parts project upward as plateaus and mountains.

much less perfect than in a liquid and of course requires a much longer time.

This assumed mechanism for preserving balance in the crust is illustrated by Fig. 267. A mountain mass is eroded deeply, and the débris is transported across an adjacent low plain, to be deposited in an ocean basin. For a time the crust can bear the strain, but eventually the mountain segment becomes abnormally light through loss of mass; the surrounding crust, and especially the part loaded by the sediments, forces deep-seated rocks to move laterally and buoy up the lightened segment. At first thought it may appear that this mechanism would make it impossible for erosion ever to wear highlands to a low level, as there would be constant rejuvenation. However, as the deep rocks are denser than those near the surface the amount of uplift cannot equal the thickness of lighter rock removed; and therefore mountains can finally be brought low, though only by long-continued erosion.

A striking confirmation of the isostatic theory is furnished by the areas in North America and Europe that were covered by great ice sheets during Pleistocene time. As the glacial ice melted the sea invaded much of the glaciated area, but was excluded by later uplift. Bones of whales and dolphins, and other evidences of this late submergence, are found in the region of Lake Champlain and Montreal, 600 to 700 feet

above present sea level; and similar evidence is found in Scandinavia. The plain inference is that the glacial ice was an overload which depressed the land. After the load was removed some time was required for restoration of balance by slow plastic flow in the rocks at depth. Although it is thousands of years since the ice disappeared, perhaps the adjustment is not yet complete, as parts of Scandinavia are still rising at the rate of 2 or 3 feet in a century. Strong tilting of the Great Lakes basins in

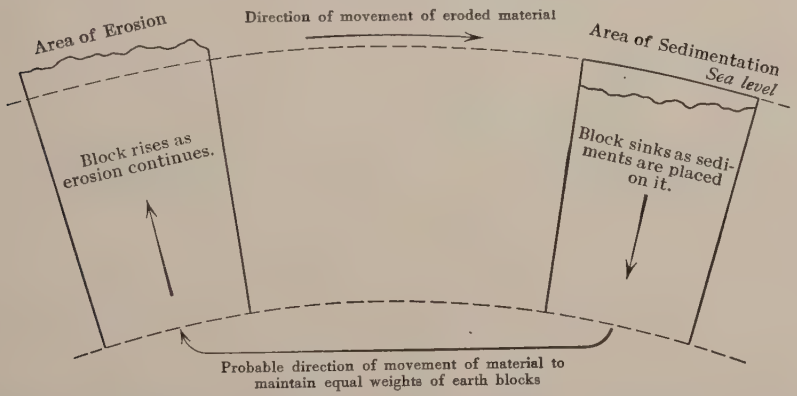
Area of Erosion
Direction of movement of eroded material
Area of Sedimentation
Sea level
Block rises as erosion continues.
Block sinks as sediments are placed on it.
Probable direction of movement of material to maintain equal weights of earth blocks

Fig. 267. — Diagram to explain the supposed mechanism to restore balance when load is shifted on the surface by erosion. The eroded area, at the left, continually loses weight, which is added to the low area at the right. In time this loaded part of the crust sinks, and forces deep-seated rock to move horizontally by slow plastic flowage. This material moves underneath the eroded "block," which rises to a position of proper balance. The plains area, across which eroded sediments are transported, suffers neither depression nor uplift. The representation is highly artificial, as there are no definite, freely moving "blocks" in the lithosphere. (Bowie.)

North America is ascribed to the postglacial uplift, which seems to have been greatest in southern and eastern Canada, where presumably the glacial load was greatest.

The general fact of isostasy appears to be established; but many uncertainties are connected with the subject. We do not know the depth at which plastic flow occurs during adjustment, though from several lines of reasoning it is argued that this depth is only a few tens of miles. The size of a load necessary to start the mechanism of adjustment is not known. These and other problems may be solved by continued study.

CONCLUSION

The problems connected with the Earth's interior are fascinating but difficult. Their solution requires the coöperation of geology, physics, mathematics, and other branches of science. Perhaps some of them are

quite insoluble; but a promising beginning has been made. It is well to keep in mind that some of the present views and conclusions are tentative, and may be changed by continued investigation. Speculations in this field are numerous, and these should be kept distinct from legitimate inference and proved fact.

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A classic paper, admirably written, in which the term *isostasy* was first proposed and defined.

CHAPTER XVI

THE ORIGIN AND HISTORY OF MOUNTAINS

Mountains are of great importance in geology, as they furnish a large part of the information on which the science is based. Dynamic processes, such as stream erosion and glaciation, are especially vigorous and their effects strikingly evident in high ranges. The uplifting of rock masses to great heights has resulted in dissection to unusual depth; and consequently a mountain region affords excellent opportunity not only for descriptive study of rock formations, but also for deciphering the history they record. But if the mountains give aid in solving many problems relating to the Earth, they also present mysteries in themselves. Why have sea floors of remote periods become the lofty highlands of today? What generates the enormous forces that bend, break, and mash the rocks in mountain zones? These questions still await satisfactory answers; but the architectural features of great ranges at least offer hints as to their origin, and are worthy of study for their own sake.

MOUNTAIN UNITS

An isolated high mass that rises above comparatively low surroundings is described simply as a *mountain* or a *peak*. Examples are Stone Mountain in Georgia, Mount Monadnock in New Hampshire, and Mount Etna in Sicily. More commonly mountain masses do not stand alone, but are parts of distinct units that vary in size and plan, from small irregular *groups*, like the La Sal Mountains of Utah and the Black Hills of South Dakota, to the enormous belt that extends more or less regularly from Gibraltar eastward to the East Indies. Descriptions of the larger units or their parts employ somewhat loosely the terms *range*, *system*, and *chain*. As it is desirable to use descriptive terms with a definite meaning, the usage proposed many years ago by J. D. Dana will be followed generally.

A *mountain range* is either a single large, complex ridge, or a series of neighboring parallel ridges that form a more or less continuous and compact unit. Excellent types are the Sierra Nevada in eastern California and the Front Range of Colorado. A group of ranges that are obviously similar in their general form, structure, and alignment, and presumably owe their origin to the same causes, constitutes a *mountain*

system. Thus the Southern Rocky Mountain system, extending from Wyoming through Colorado into New Mexico, is made up of the Sangre de Cristo, Front, Sawatch, and other great ranges formed in the same geologic period. A different designation is needed for a mountain belt such as the Appalachians, which includes a number of groups and ranges diverse in plan, structure, and geologic date. Accordingly all of the mountains in the broad belt extending from Alabama to Nova Scotia and Newfoundland are grouped together as the *Appalachian chain.* The Rocky Mountain chain includes both the Southern and the Northern Rocky Mountain systems, constituting a great unit that extends from near the Mexican boundary through the United States and western Canada.

But a still more comprehensive term is needed in referring to a series of chains, systems, and ranges that make a more or less compact belt of large extent. Following the famous traveler Humboldt, a Spanish word has been borrowed for this purpose. All of the mountain units in western North America, from the eastern border of the Rocky Mountains to the Pacific coast, are known as the North American *cordillera.* Similarly the entire broad mountain belt that extends almost continuously from Alaska to Cape Horn is known as the American *cordilleras.* However, the same term has not been adopted universally for the major mountain belts of the Earth. In the literature the great mountain unit of southern Europe and Asia is designated variously, as the *Mediterranean* (or *Eurasiatic*) *chains, zone,* or *belt.*

ORIGIN OF MOUNTAINS

Mountains owe their origin to different agencies, and differences in the structure and the plan of mountain units are due largely to this fact. Ultimate causes of mountain building are in large part obscure; but regardless of this fact the principal agencies involved may be stated as differential erosion, volcanic activity, and movements of the crust. Commonly two or more processes combine to produce complex results. However, the discussion will be clarified by a classification that recognizes the dominant agencies.

Residual Mountains

High plateaus suffer differential erosion, and during a late stage in the process some of the more favored residuals may have sufficient height, in relation to their surroundings, to be called mountains. Some of the larger buttes in western United States are examples. The Catskill Mountains in New York represent remnants of an extensive high plateau,

the greater part of which has been removed by stream erosion. An early stage in a similar development may be seen in the Grand Canyon of the Colorado, where a number of pyramidal erosion remnants rise from the depths of the chasm (Fig. 268). These are dwarfed by their surroundings; but if they could be placed on a plain they would rise to mountain heights. It is not difficult to imagine that in a later geologic epoch the present youthful plateau will have been dissected thoroughly by the

Fig. 268. — Vishnu's Temple, Grand Canyon, Arizona. Mountain-like remnants, left during dissection of Colorado Plateau by Colorado River and tributaries. Part of the level, undissected plateau, many miles distant, appears in background. (U. S. Geological Survey.)

Colorado and its tributaries, and will be represented only by scattered residual mountains separated by plains and valleys.

Plateau blocks from which residual mountains are derived may consist of horizontal rock formations, of homogeneous crystalline rocks, or of folded and faulted strata that were peneplaned by previous erosion. Obviously the resulting mountain forms will be influenced by the original structure. Residual mountains are also known as "mountains of erosion." In a strict classification, however, it must be considered that deep differential erosion is made possible only by great uplift, and therefore movement of the crust is an important factor in producing residual mountains. As plateaus, in their initial form, are entirely distinct from mountains, a rigid classification might insist that the erosion residuals be

called "plateau remnants" and not mountains. However, this would be an academic distinction without practical value.

Volcanic Mountains

Some of the loftiest peaks in the world, such as Chimborazo (20,517 ft.) and Aconcagua (23,393 ft.) in the Andes, and Kilimanjaro (19,321 ft.) in Africa, have been built directly by volcanic action. Many such peaks, however, have their bases on high plateaus, and therefore their true height is much less than the altitudes above sea level indicate. A large number of volcanic islands are great volcanic piles, and some of them appear to be seated directly on the deep ocean bottoms. Thus the island of Hawaii, measured from the Pacific floor to the highest peaks, has a total height of about 30,000 feet; and the entire mass, so far as can be judged, consists of extrusive basaltic rocks.

Volcanic peaks commonly are superimposed on mountains that resulted from other causes. (All mountain masses formed directly by igneous extrusion are sometimes called "mountains of accumulation.") In areas of widespread volcanism the volcanic materials build up extensive plateaus, such as the Absaroka Plateau east of Yellowstone Park, and the great Columbia Plateau of Washington and Oregon. From units of this kind residual mountains are produced in time by differential erosion.

Mountains Whose Structure Reflects Crustal Movements

The mountain types considered above are important geologically, as they form groups of considerable size in the world today and have had wide distribution during past geologic periods. Differential erosion is one of the major factors in producing mountain relief. However, this factor could not operate unless large areas of the continents were raised high above sea level. Furthermore in most of the dominating mountain units the relief has been conditioned, either directly or indirectly, by localized movements which have caused more or less severe disturbance of the rocks. In some of the young mountain systems much of the local relief was caused directly by these movements; in older belts, which have experienced deep denudation and perhaps more than one rejuvenation by regional upwarping, the original disturbance of the rocks may be important chiefly in guiding erosion. But even if the present mountain relief is due chiefly to differential erosion, any characteristic structure produced by crustal movement has large importance in classifying mountain units. According to the nature of the movements, as indicated in the resulting structure and form, mountains of this general type may be

divided into four classes. (1) Dislocation or faulting on a large scale results in relative uplift of rock masses, with or without tilting. Ranges whose structure is produced chiefly by this process are *fault mountains*. (2) Some vertical movements result in arching of the rocks into a general domal form, either nearly circular or somewhat elliptical in plan. *Dome mountains* result from this process. (3) More commonly the forces deforming the crust produce large plications, or parallel anticlines and synclines, giving rise to the structure of *fold mountains*. (4) In most of the great mountain belts we see the combined effects of two or more types of movement, particularly folding and faulting, with complications produced by igneous intrusion. The resulting mountains are of the *complex* type, although locally they may be classified according to the process that has played a dominant rôle.

Excellent examples of each mountain type exist; but Nature is complex, and consequently various combinations of the different types are most common. The influence of erosion in varying degree is evident in all mountains, regardless of type.

Fault Mountains

Assume that a system of intersecting fractures, reaching to great depth, divide part of the Earth's crust into blocks or masses of very large dimensions. Mountains may be produced directly by movements of several kinds. (*a*) If the region is initially a high plateau, some of the blocks may be depressed several thousand feet, leaving other blocks in their original positions, to form mountain ranges. (*b*) Regardless of original altitudes, some of the blocks may be elevated to mountain heights, by forces acting largely in the vertical direction, leaving adjacent blocks relatively depressed. (*c*) All of the blocks may move downward or upward, but differentially, so that in the end some stand much higher than others. (*d*) Each of several blocks may be tilted or rotated, one edge being elevated and the opposite edge depressed. The faults on which movement occurs may be either normal or reverse. Regardless of the nature of movement, a series of neighboring ranges obviously due to faulting may be called either *fault mountains* or *block mountains* (Fig. 269).

Fault blocks as we actually see them have been more or less modified by erosion. Débris worn from the high masses tends to bury those at low elevation. In time this combination of erosion and deposition may nearly or quite obliterate the mountain relief, especially in a region of interior drainage where all the débris is retained. At a later date, as a result of broad regional uplift perhaps accompanied by change of climate,

the original pattern of ranges and intermontane troughs may be etched out by vigorous erosion. Obviously these resurrected ranges are the direct result of differential erosion, and strictly they are residual mountains. However, it is recognized that the original faulting was a strong factor in guiding denudation and continues to be reflected in the mountain forms. Accordingly the original structure is given large emphasis,

Fig. 269. — An example of fault mountains. A great fault zone, somewhat irregular, marks the base of the range. Downthrown block, at left, covered with lake deposits. Scarp 2000 feet high. Bluejoint Rim, Oregon. (Airplane view by R. E. Fuller.)

and it is common to use the designation *fault mountains* or *block mountains* even for units that have been greatly modified by erosion.

The Sierra Nevada of California is a tilted crust block 400 miles in length and approximately 100 miles wide. Its eastern edge has been uplifted two miles or more, to form an abrupt eastward-facing scarp. Roads from the east ascend this precipitous front with difficulty; but west of the crest they descend on a long, gentle slope — the tilted upper surface of the block. In the Great Valley of California, sediments thousands of feet deep have accumulated on the depressed portion of the rotated mass. A great series of fault mountains lies in the Great Basin of Nevada and neighboring states. Ranges in that region are so similar in character that they are known collectively as the Basin and

Range System. The great dislocations responsible for these ranges and for the Sierra Nevada do not represent the first disturbance of the region. In earlier geologic periods the thick sedimentary rocks in Nevada and eastern California were folded and crushed, and great igneous masses were intruded into them. Mountains that existed soon after those ancient events disappeared long ago, and in fairly recent time the old deformed crust has been broken by great faults, to form the present generation of mountains. Some of the ranges are still growing; for within historic time movements have occurred on several of the faults, giving rise to violent earthquakes. As a large part of the Basin and Range region has no drainage to the sea, the mountains are partly buried by accumulations of their own débris.

Fault mountains in various stages of destruction are found in parts of eastern and northern Africa, in Arabia, and in central Asia. The Triassic sandstones of Connecticut and Massachusetts are broken by great faults, and the resulting blocks have a strong tilt eastward; but the mountains that were formed by these dislocations have disappeared through erosion. Similar structural relics of ancient fault mountains, representing various geologic periods, are widely distributed in all continents.

Dome Mountains

Mountains whose structure reflects crustal uplift of distinctly domal character may be classed together, regardless of size or the exact cause of the uplift. The simplest and best understood are laccolithic domes, made by the bowing up of strata above thick, lens-shaped intrusions of liquid rock. Ordinarily a dome of this kind that is high enough to be called a mountain has lost more or less of its original cover through erosion; and not uncommonly the resistant igneous mass, almost completely denuded, stands within circular or elliptical ridges formed by the upturned edges of the more resistant strata. An excellent example is Bear Butte, one of many laccolithic mountains in the vicinity of the Black Hills (Fig. 190). The Henry Mountains of Utah, classic examples of the type, are a large group of laccoliths in various stages of denudation. But not all mountains of this kind have ideally simple structure. The intruding magma not uncommonly ruptures the covering strata and lifts them irregularly, as in the Moccasin Mountains of Montana. A compound laccolith, such as the Mount Holmes mass in western Colorado, presents, after some erosion, a confused arrangement of the igneous rock and the intruded strata.

Dome mountains on a larger scale are illustrated by the Black Hills of South Dakota. In a casual journey through these mountains the

traveler may gain an impression of disordered arrangement; but a good map or a block diagram of the entire unit reveals a beautiful symmetry of structure, involving an elliptical area about 100 miles in length by 50 in width (Fig. 270). In the middle and eastern portions the uplifted sedimentary formations have been stripped away, and there the crystalline basement rocks have been carved into ridges and peaks, including Harney Peak (7242 feet above sea level), the highest point east of the Rockies. Around the flanks of the uplift the upturned edges of sedimentary formations, yielding to erosion at different rates, make alter-

Fig. 270. — Diagram of the Black Hills uplift, South Dakota. View looking north, along the longer axis of the elliptical dome. The wide valley encircling the dome is commonly known as "The Race Track." (Newton, U. S. Geog. and Geol. Survey.)

nating high and low belts which encircle the uplands. Some of the heavy limestone members still cover the western half of the uplift, forming the "limestone plateau." A careful profile and section, in which the eroded formations are restored to their original positions, shows that the dome is steeper on the eastern than on the western flank, and that the greatest erosion has occurred in the area of maximum uplift. The restored section indicates that the top of the dome was elevated 9000 feet, although the present height above the surrounding plains is less than 4000 feet.

In contrast with a laccolith, the crystalline rocks exposed at the core of the Black Hills were not intruded as magmas, but are much older than the domed sedimentary formations. Therefore the force by which the uplift came about was applied much deeper than any exposed part of the

crust. If the cause of uplift was the rise of igneous material, the molten mass was very large, quite symmetrical in form, and very deep-seated. The connection of some igneous activity with the movement is indicated by numerous laccoliths, which form small satellitic domes in the northern part of the Hills. Whatever may have been the part performed by igneous magmas in causing the main uplift, it is very probable that the horizontal pressure by which the Rocky Mountains were folded was an important factor also in shaping the Black Hills dome. The long axis of the Hills is parallel to the Rocky Mountain front, and the eastern side of the dome is especially steep, as if there had been strong thrust from the west. According to this view the bulging up of the Black Hills was merely a local incident in forming the Rocky Mountain structure; and the laccoliths superimposed on the larger dome represent still smaller incidents in the general process.

It is evident, on reflection, that the Black Hills uplift did not become an actual mountain group without the work of erosion. If the youngest strata involved in the movement still extended unbroken across the summit, forming a broad, smooth dome, the area would be a small plateau rather than a mountain. It is essential, therefore, to keep in mind the limitations of any scheme of classifying mountains. The part played by erosion is of great importance in connection with every other process. However, the crustal deformation which controls the relief forms is important enough to merit recognition in a classification. The limit of size separating dome mountains from plateaus must be somewhat arbitrary; but as the Black Hills uplift is a definite unit of moderate size, with strongly defined boundaries, it seems proper to emphasize the structural form, provided the various steps in fashioning the highlands are kept clearly in mind.

If the Black Hills should be worn down to a peneplain by prolonged denudation, and a later warping movement should reëlevate the Great Plains region several thousand feet, subsequent erosion would be guided by the old structure and a group of residual mountains similar to the present Hills would be produced eventually.

Fold and Complex Mountains

Mountains in which the rocks are strongly folded and broken are commonly described according to their internal structure, regardless of the later chapters in their history. Some old mountain units may be strictly remnants of erosion, and therefore residual mountains. However, if they give evidence that a certain type of crustal deformation attended their early development, these structural characteristics are

used as the basis of their classification. For example, the Appalachians have had a long and varied career. The original chain suffered erosion through long ages, and almost or quite disappeared; and the present Appalachian ridges have been etched out after a later upwarping of the area containing the old mountain "roots." Nevertheless, remnants of the original structure are clearly visible, and are recognized as an important feature of the mountain zone. It is well to keep this example in mind throughout the discussion that follows. Deformed rocks are characteristic of all great mountain zones, and the development of this deformed structure will be given prominent attention in outlining mountain history; but it is not to be taken for granted that the deformation gave rise to the present mountain elevations. Mountain *structure* and mountain *elevation* may not have any direct relation to each other. Nevertheless the structure continues as a dominant factor in determining relief because it guides differential erosion.

All the great mountain chains of the Earth include folded sedimentary rocks as a conspicuous part of their structure. These chains, therefore, are sometimes classed together as *fold* mountains, although faulting, igneous intrusion, and other important processes besides simple folding have played some rôle in their origin. Strictly speaking, every great system is more or less *complex* in its structure; but certain mountain units exhibit fairly regular plication of rock formations as their outstanding structural characteristic. The Jura Mountains in Switzerland and parts of the Appalachian Mountains in North America are excellent examples. The Rocky Mountains and the Alps are characterized by enormous thrust faults in addition to folds, and consequently are illustrations of complex units. However some parts of the Appalachians also are complicated by thrust faulting; and as there are all stages of gradation between the simpler sections of this chain and the almost incredible complexity of the Alps, it is clear that *fold* and *complex* mountains cannot be separated as sharply contrasted structural types. Therefore it is desirable to include mountains of these two classes in a unit discussion; although the treatment logically emphasizes the simpler folding first, and then proceeds to more complex processes and results.

Considering the Earth as a whole, the finest exhibitions of geologic phenomena are furnished by the mountains with folded and complex structure. There are to be found the upturned and dissected strata whose kinds, thickness, included fossils, and structure furnish the most effective key to past events. Commonly the making of such mountains has been accompanied by igneous activity, and the sections now exposed reveal both intrusive and extrusive masses of various types. Many of the most important ore deposits occur in these zones of disturbance, both

recent and ancient. Some of the youngest complex ranges are the theaters in which many agents of erosion, as well as crustal movements and volcanism, play their most active rôles at the present time. For many reasons, therefore, the great mountain belts merit special consideration.

General Characteristics of Fold and Complex Mountains. — From examination of a globe or a world map it is apparent that the prominent mountain belts are elongated generally parallel to the continental margins. This relation is especially striking in the American cordilleras, the Appalachians, the Scandinavian chain, and the great Eurasiatic mountain zone. Each major belt is composed of numerous ranges disposed somewhat irregularly but with the same general orientation. Some of the ranges are nearly straight in plan; but many are strongly curved into the form of great bows or arcs. The Alps, Carpathians, and Himalayas are striking examples of this arcuate type.

Generally the exposed portion of each range is made up in part or wholly of distorted sedimentary formations. Commonly these strata, now on the flanks or even on the highest summits of the ranges, represent deposits in former seas or on deltas and in marshes bordering the sea. Owing to the strong folding and faulting of these strata, followed by planation and dissection through erosion, the full thickness of the sedimentary cover is exposed in many places. In some mountain belts these thicknesses are astonishing; 4, 5, or even 6 miles are by no means exceptional values, and in some mountain areas the total sedimentary sections exceed 40,000 feet. It will occur to some readers that similar thicknesses may be common also outside of mountain zones, but are not known because conditions favorable for their revelation do not exist. However, natural exposures and well records indicate clearly that sedimentary formations grow conspicuously thinner away from a folded mountain belt. Thus the strata in the Appalachians average 20,000 feet or more in thickness along the central axis of the folded tract; but at no great distance to the west the thickness is less than 10,000 feet, and in the Mississippi Valley it is only 4000 to 5000 feet. On the east side of the Appalachians the sedimentary strata do not exist, and it will be shown presently that the deposits never extended far eastward from their present limit. Therefore the excessively thick sediments occupy a long and relatively narrow belt that corresponds closely to the axis of the folded chain. This general relationship exists also in the Rockies, the Andes, the Mediterranean ranges, and other great mountain systems. It is a natural conclusion that the accumulation of abnormally thick sediments had a significant connection with the development of each of these mountain units.

The typical history of ranges with fold structure falls into three general

divisions: the *preliminary stages* in which certain processes prepare the place and some of the material for the future mountains; the *period of crustal movements*, in which the folds and related structural features are produced and the initial uplift occurs; and *subsequent stages*, during which the mountains experience various modifications through erosion and repeated vertical movements. This subdivision of the history is useful, if it is understood that the periods are not sharply separated, and that some of the processes involved operate simultaneously. Thus some incidents of folding and thrusting occur long before the accumulation of sediments is complete; and inevitably the modifying influence of erosion dates from an early stage of uplift, while the orogenic or mountain-making processes are active. The growth and decline of every range is the result of a slow, complicated interplay of forces that act either continuously or recurrently during a very long geologic interval. Recognition of general stages in the history serves to clarify a brief discussion; but these stages overlap and merge irregularly into one another, so that the whole sequence is typified by the cross profile of a mountain chain itself, in that it rises gradually and with interruptions to a culmination and declines in the same way.

Preliminary Stages: Development of Geosynclines. — Sedimentary strata in the great ranges consist of conglomerates and sandstones mingled with shales and limestones. It is clear that thick deposits containing coarse sediments must have been laid down near the margin of a land that suffered prolonged erosion. The great thickness of deposits may suggest that sedimentation began in an excessively deep basin. However, there is unquestionable evidence that nearly all sediments involved in mountain folds were laid down in shallow water or at only moderate depths. Accumulation of such deposits to a total thickness of several miles indicates that slow subsidence of the sea floor was continuous or recurrent while deposition was in progress. Moreover, as enormous volumes of coarse sediments were delivered into the subsiding basin repeatedly, the wasting land must have risen continuously or recurrently adjacent to the area of sedimentation. In any case, the preliminary structure that determines the location of a future range appears to be a sinking trough into which the waste from near-by land accumulates to unusual depth. An elongated subsiding tract of this nature is known as a *geosyncline*. Modern examples may be the Great Valley of California and the enormous Indo-Gangetic flood plain of India.

In the Appalachians various features of the strata indicate that conditions within the old geosynclinal trough fluctuated repeatedly. Sandstones and shales with abundant ripple marks and mud cracks are interbedded with thick limestones that contain marine fossils. Such

relations imply a shifting coast and considerable variation in depth of water. In fact at some periods the sea gave place to great delta plains or to enormous swamps in which materials for coal beds accumulated. These changes depended on the relative rates of subsidence and sedimentation. If sinking of the trough halted for a considerable time, ac-

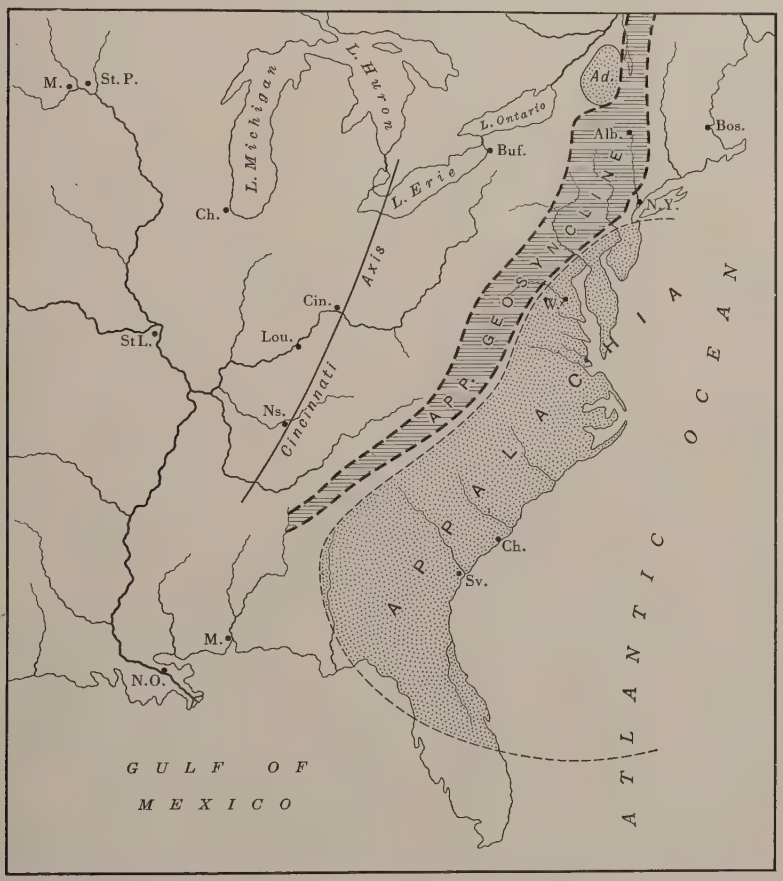

Fig. 271. — Map showing the situation of the Appalachian geosyncline and of the old land Appalachia. *Ad*, mass of the Adirondacks.

cumulating sediments made the sea shallow or even displaced the sea water entirely over wide areas. With renewal of subsidence the water came back. If the adjacent land was elevated rapidly for a time, erosion may have been stimulated sufficiently to keep the seaway full even though subsidence of the trough was continuous.

From study of the sedimentary sections in the Appalachians it is clear

that the coarser sediments are on the east, and to the west these give way to marine shales and limestones. Therefore the land from which sediments were eroded lay to the east of the geosyncline. A narrow belt of ancient rocks near the present coast presumably represents the western edge of the former land; but it must have extended far eastward, over the area of the present continental shelf, in order to have volume adequate to explain the vast quantities of sediments it supplied. The name "Appalachia" is applied to this ancient land of unknown extent (Fig. 271). The sea that lay west of it, covering much of the present Mississippi Valley region, was shallow and fluctuating.

Similar histories have preceded the making of other great ranges. The Rocky Mountains grew up from a great geosyncline that stretched from the Gulf of Mexico to the Arctic, with highlands to the west. The old land that furnished sediments for the Alpine geosyncline lay to the north of the Alps. Strata folded in the Caucasus were derived from lands to the south while seas stretched northward over Russia. Thus it may be accepted as a general principle that on one side of the mountain zone lies an area of much older rocks, the source of the folded sedimentary deposits. The time occupied in the accumulation of sediments in the geosyncline extends over long geologic periods.

Period of Crustal Movements.—The period of relatively quiet preparation, of long-continued erosion and sedimentation and slow movements of land surface and sea bottom, gives way to a period of greater activity in which the Earth's outer shell yields to powerful lateral pressure. By this pressure the accumulated load of sediments is thrown into folds, crushed and mashed together into the disordered arrangement characteristic of the great chains. The process and its results thus simply stated are in reality very complicated, with different phases and with divergent features in different regions, some of the more important of which demand separate consideration. We shall take up first the operating forces and then the results produced.

It is clearly evident, from the structural features found in complex mountains, that crushing of the old geosyncline and its burden of sediments was performed by forces acting in a lateral direction, tangential to the Earth's surface. Thus in zones of most intensive folding, the folds not only become closed so that their limbs are in contact, but they are even more severely compressed, with mashing of the beds and the production of very complicated structures. This is illustrated by sections in the Alps and in the Appalachians (Fig. 272). Considering the scale of these sections it is impossible to imagine the formation of such structures except by transverse compression of great magnitude.

The varied phenomena of folding shown in the mountains may be

imitated by lateral compression applied to a sequence of artificial strata composed of some plastic substance, such as wax or clay, placed upon one another. If some of the layers are made somewhat harder and stronger than others, and the series is laid in a firm trough or box, one end of which may be forced inward by turning a screw, the stronger layers tend to form large and simple folds, whereas the weaker members are distorted in a more complicated way. With continued pressure the folds may be overturned and broken. The displacements and dislocations, the folding and faulting of the strata, produced in miniature by this

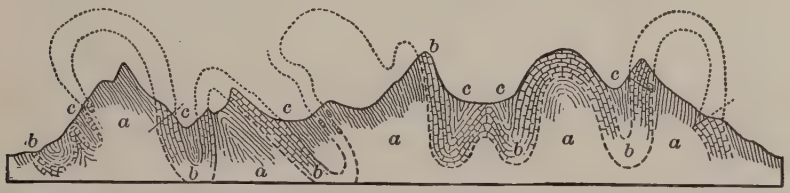

Fig. 272. — Section across the Säntis Alps, N. E. Switzerland (after Heim, somewhat modified). *a*, shales, breccias; *b*, massive limestone; *c*, shales, thin limestones.

method, are similar to those observed on a great scale in the mountain ranges (Fig. 273).

Cleavage. — It has been shown that cleavage in metamorphic rocks, such as slates, is produced by great pressure, and that the planes of cleavage are developed at right angles to the direction of pressure. Many of the rocks of the great ranges in the zone of intensive folding have been turned into gneisses, schists, or slates, depending on their original composition and other factors. This alteration becomes more evident as the inner portions of the compressed masses are exposed by erosion. Observation shows that the planes of cleavage usually stand at high angles and not uncommonly are vertical, whereas the strike of the cleavage planes is generally parallel to the axis of the range. The direction of the compressive force, thus indicated by the cleavage, is the same as that shown by the folding.

Faulting. — It is obvious that such extreme folding of rocks could not take place without rupturing, breaking, and displacement of the strata. We find, accordingly, that the phenomenon of faulting is very common in mountain ranges. As we pass from consideration of the simpler fold ranges to those of more complex types the faulting becomes more pronounced until finally it culminates in thrust faults of enormous magnitude. The small angle of incidence of the thrust planes to the horizontal and their trends parallel to the axes of the ranges are indicative of the lateral force, or approximately horizontal compression, that has produced them.

In summation, then, we may accept it as a well-grounded fact that the structure of the complex ranges has been made by the lateral shoving, or squeezing together, of the stratified beds laid down in geosynclines.

Amount of Compression. — The magnitude of the forces involved and of the masses operated upon is indicated by the amount of compression

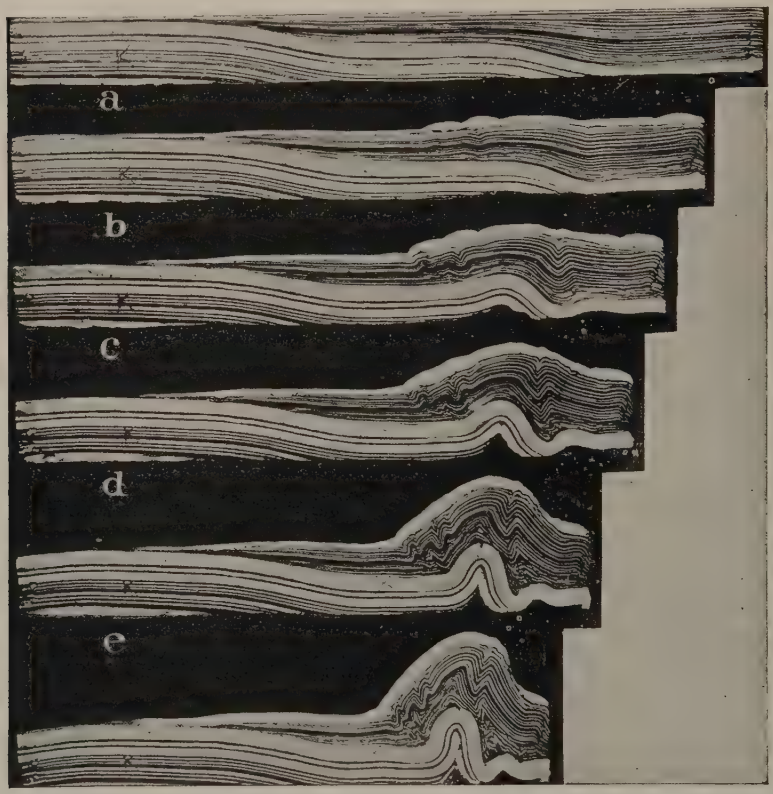

Fig. 273. — Layers of wax and plaster folded by lateral pressure, imitating structures found in mountain ranges. The thrust is from the right, and in successive layers from *a* to *e* the amount of shortening can be seen. (Willis, U. S. Geol. Surv.)

which investigation shows has actually occurred in some of the great folded belts. In the Appalachians, estimates of 40 to 50 miles, and in some sections even more, are given for the distances the original width of the strata in the geosyncline has been decreased by the mashing together of the mass. In other words, if the folded strata in Pennsylvania, which resemble a crumpled blanket, could be smoothed out toward the southeast, their extent would be increased sufficiently to cover the state of New Jersey. The Rocky Mountain structures represent a comparable

amount of shortening. Thus the original breadth of the geosynclines has been diminished by tens of miles, and in the zones of intensive folding and mashing the reduction has been one half or more.

Influence of Resistant Elements. — The old upland along whose margin the sediments have been deposited forms a resistant element in the architecture of the outer shell. It tends to rise as the geosyncline sinks, and as it becomes eroded the stronger massive rocks, igneous and metamorphic, of which its lower levels are composed, are exposed at the surface. Thus the old land becomes steadily more massive and resistant, a more unyielding block or element in the shell. By contrast the sinking zone of accumulating sediment is one of weakness; the sinking is primarily the cause for the accumulation of the sediments, but probably the growing load of deposits causes additional subsidence. Finally, when the shell yields to lateral compression, the weak sediments are crushed against the strong adjoining mass and are crumpled. Rocks in the resistant element are mashed somewhat, but there are no stratified formations in it to yield by folding. Irregularities in the margin of the old land appear to have an influence on the trend of the mountain folds, causing prominent bends or arcs. Therefore the general plan of fold ranges seems to be determined in an important degree by the situation of the old lands, which act as resisting buttresses at the time the folds are formed. Thus the Appalachians from southern New York to Alabama imitate roughly, in their sinuous trend, the former western coast of the old land Appalachia. It is suggested also that the curving trend of the Alps has been determined in large measure by old land masses, parts of which are now visible in central France, the Vosges, the Black Forest, and Bohemia.

Fig. 274. — Diagrammatic section across part of the Jura Range, showing simple structure and symmetrical folding.

Variations of Folded Structure. — It is to be expected that the results of folding should differ considerably in character between separate ranges, or between distant parts of the same range. Differences in thickness of sediments, in proportions between strong and weak formations, in the form of old rigid masses that transmitted the thrust, and in severity of the lateral forces, are reflected in the individuality of mountain folds. The Jura Mountains, a small member of the Alpine system, furnish classic examples of symmetrical, upright folds (Fig. 274). These folds were produced far out in front of the Alps proper, in a relatively

thin sheet of strata, as an incidental effect of the forces that deformed the greater Alpine zone. The Appalachians present a wider variety of fold structures. In the slate and anthracite regions of eastern Pennsylvania the folds are closely compressed, many of them to the isoclinal stage, and the axial planes are strongly overturned toward the northwest. Farther west in the state the folds tend to be open and upright; and the deformation dies out gradually westward. Going to the south, through Virginia and Tennessee and into Alabama, we find that many of the folds were ruptured by the severe compression and developed into thrust faults (Fig. 275). This kind of complexity is especially pronounced in the Alps, which merit special description.

Thrust Faults and Recumbent Folds of the Alps. — Alpine structure is characterized by great folds that have been pushed over to a horizontal attitude, and by flat thrusts that are related to these overturned folds. These features are developed on an unprecedented scale, with the result

Fig. 275. — Section 12 miles long illustrating Appalachian structure near Greeneville, Tenn. (Slightly modified from Keith and Willis.)

that the Alps consist of a series of great rock sheets, driven one over another and overlapping like the shingles on a roof. The Germans call the individual sheets *decken;* the French refer to them as *nappes.*

Because of their location, the Alps have received more intensive study than any other mountains. Accordingly, in spite of astonishing complexity, their structure and history are well known. Like the Appalachians, they resulted from deformation of thick marine deposits; but a large part of the Alpine sediments bears evidence of deposition in deep water, far from any shore. Land lay to the north, in the present position of central Europe, where mountains of nearly the same date as the Appalachians were being eroded. Orogenic movement began in Mesozoic time, with pressure from the direction of Africa. The soft sediments on the sea floor were bowed up slowly, until islands, and chains of islands, appeared above sea level. During early Tertiary time the compression accelerated powerfully, and an enormous rock sheet was driven northward over Europe. Beneath this sheet the plastic sediments suffered extreme distortion. With recurrent thrusting during the Tertiary other sheets were driven forward, and all were severely folded (Fig. 276). Erosion cut valleys and "windows" through the sheets, exposing the entire series; and in parts of the Alpine area nearly the whole of one or more sheets has been swept away, leaving remnants of old rocks to form

isolated peaks standing on younger rocks that were overridden and covered during the thrusting movement. Isolated peaks that have this

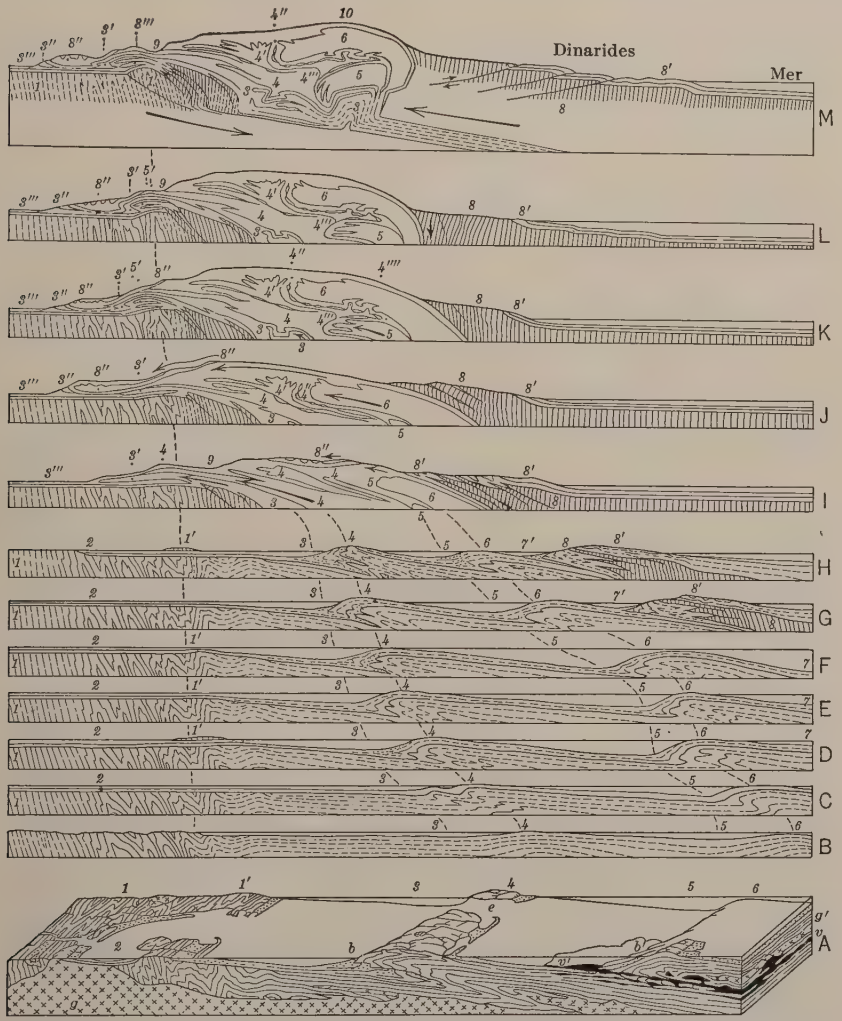

Fig. 276. — Development of Alpine structure. *A*, block diagram representing part of the Alps in an early stage. *B* to *M*, cross sections to show successive stages from beginning of folding (*B*) to the final intricate structure of thrusts and folds (*M*). Figures show horizontal shifting of corresponding masses. Northwest on the left. (Emile Argand.)

anomalous relation are called "mountains without roots." The Matterhorn and the Mythen are famous examples. Some of these masses are 50 or even 100 miles north of their original positions. Heim, the

great Swiss master of Alpine structure, tells us that the Alpine zone as a whole was made narrower by considerably over 100 miles due to the thrusting and folding. Locally, as in the Simplon tunnel section, the original width was reduced as much as 90 per cent.

Mountain Elevation. — There is a natural tendency to assume that mountains rise to greater heights continuously with folding and thrusting, as a logical result of crowding excess material into a narrow zone. For a long time, indeed, this conclusion was taken for granted, and attempts were made to compute the original height of eroded ranges by determining the amount of shortening due to folding. As the steps in mountain history become clearer, however, it is found that much of the actual elevation occurred at a distinctly later time than the folding and thrusting. After the Rocky Mountain deformation in early Tertiary time, the folded and faulted area was eroded to a nearly even surface at a low altitude; and the present great heights in the Rockies are due to vertical uplift in late Tertiary time. Similarly, after much of the thrusting and folding was complete the Alps had only moderate height, and the sea washed the flanks of the range both on the north and on the south. In very recent geologic time a movement of vertical uplift carried the Alpine summits to great height. The Andes and the Himalayas have had a similar history.

If the principle of isostasy is kept in mind, it does not seem strange that horizontal pressure acting alone fails to force the folded zones to great height. An overload results from horizontal transfer and piling up of the rocks, and the mountain area continues to subside under the increasing weight, in order to maintain approximate balance with adjacent parts of the crust. Therefore even the enormous amount of material heaped together in the Alpine zone did not result at once in high mountains. Presumably there was slow flowage outward from the deformed area, in a deep plastic zone, to prevent an extreme overload. The cause of later vertical uplift is a matter for speculation. It is suggested that the cold rocks of the upper crust, carried to a deeper and hotter zone by the continued heaping up and sinking, slowly changed their state due to heating and expanded. Such increase in volume would not change the total weight of the mountain mass, and therefore the surface would rise without disturbing isostatic balance. In a qualitative way this explanation is satisfactory; but the problem involves many unknown factors and therefore cannot be solved quantitatively.

Rôle of Igneous Agencies. — Although the making of mountain structures by compression appears to be independent of direct igneous action, and some ranges contain little or no visible igneous rock, nevertheless an upwelling of magma to produce both intrusive and extrusive bodies has

been a common incident in mountain making. The effect of this action is to modify the structure due to folding and faulting, and by the addition of large massive bodies to increase the rigidity and strength of the mountain zone. Probably the most effective way in which this happens is by the intrusion of great batholiths, usually composed of granite, into

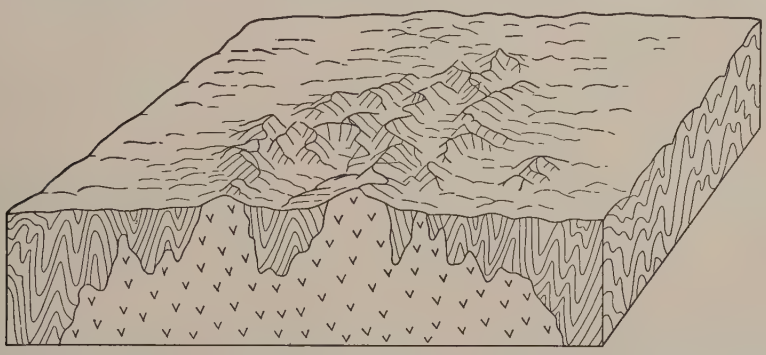

Fig. 277. — Diagram to represent the granitic core of a mountain range after prolonged erosion. The granite batholith cuts across folded and metamorphosed sediments.

the inner, lower portion of the range. A granite intrusion of this nature may become exposed later by deep erosion, and is then spoken of as the "granite core" of the range (Fig. 277). Intrusion of the heated magma, combined with the folding and mashing of the strata, causes profound metamorphic effects over wide areas. As great intrusive bodies of this kind generally cut across the folds and thrusts, it appears that the intrusion occurred late in the period of orogeny and had no direct con-

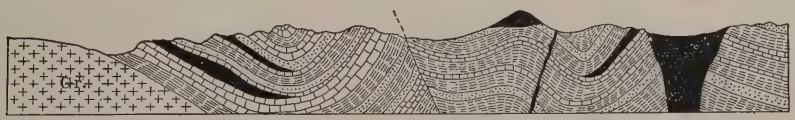

Fig. 278. — Section illustrating intrusions of igneous rock (black) in a folded and dislocated mountain region. *Gr*, edge of a granite batholith.

nection with the compressive forces. Possibly some of the batholiths were made by melting of rocks forced into the depths during the folding and consequent sinking. Invading granitic masses of this character are conspicuous features of the Coast Range in western Canada, of the Sierra Nevada, the Green and White mountains in New England, the Caucasus, and the mountains of Scotland and Norway.

Intrusions of molten magmas may not only make great batholiths, but pressing upward along belts of weakness caused by folding and faulting, they may form intrusive sheets, laccoliths, dikes, and other bodies (Fig. 278); or, attaining the surface, they may be extruded as lava flows or

give rise to violent volcanic action. During the formation of the Rocky Mountains, Wyoming and Montana were the scene of great volcanic activity, the dying phases of which are still evident in Yellowstone Park. As a result of the folding and faulting of strata, and the intrusion and extrusion of magmas, there were produced ranges with geologic structures of wonderful complexity, now revealed to us by deep dissection.

Subsequent Stages. — Although mountains may be classified according to the crustal movements involved in their formation, it is not to these processes alone that the mountains in their present form are due. Hand in hand with uplift of the masses goes the work of erosion, that mighty chisel of Nature, which shapes and carves them into the forms familiar to us. In a sense, therefore, all mountains are *residual*, as erosion starts with the beginning of uplift, proceeds during the whole period of orogenic activity, and continues so long as the highlands exist. During the subsequent stages the history of a range consists chiefly of progressive changes due to erosion, by which the mountains reach maturity and old age.

Crustal movements in the mountain zone do not cease entirely, however, at the time the ranges reach their maximum height. As erosion proceeds its work is partly offset by recurrent upwarping of the region. Old mountains rejuvenated by a strong uplift are sometimes called *posthumous* mountains. The old Appalachians have been bowed up repeatedly, even in recent geologic time. After their birth during the Permian they suffered erosion through long geologic periods; and although uplifts probably occurred, the entire folded tract finally was reduced to a peneplain. During the Tertiary the region was warped up strongly and dissected; and therefore the present mountain ridges are strictly residual mountains. The latest pulses of uplift are recorded by high terraces along the streams.

This tendency of mountain ranges to renew their growth may be connected with isostasy. After a large volume of eroded débris is transferred from the mountains to the plains or the ocean basins, the mountain mass is forced to rise to restore equilibrium. It does not rise to its original height, however, probably because the plastic rock forced in at the base of the rising mass has a higher density than the rock eroded from the surface. In time, therefore, by removal of an enormous volume of material, the mountains may be reduced permanently to a low level.

Geologic Date of Mountain Making

The geologic period in which a geosyncline has been crushed, and the strata compressed into mountain folds, is fixed by determining the age of the latest strata involved in the folding, and of the oldest undeformed

beds that lie upon the disturbed rocks about the mountain flanks. Obviously the folding is younger than any of the folded or disturbed beds and older than any that are undisturbed (Fig. 279). The closeness of dating by this method evidently depends upon the length of the interval between the dates of deposition of the two sets of strata. Thus if the youngest folded rocks are of late Triassic age, and the oldest undeformed sediments were deposited early in the Tertiary, the folding may have occurred either in the Jurassic or in the Cretaceous period. The Appalachian folding is dated rather closely, as Permian rocks are affected and Triassic rocks are not.

As many mountains have been formed not in one, but in several periods of compression, the method just explained can be used to date only the latest movement. However the earlier disturbances may be recorded

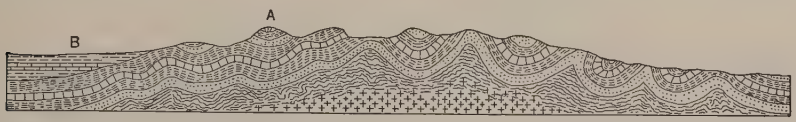

Fig. 279. — Illustrating determination of the date at which deformation in a mountain zone occurs. *A* are the youngest surviving strata that were involved in the folding; the folding is younger than these. *B* are not concerned in the process, and hence the folding is older than they are.

by unconformities. Thus if a surface of unconformity cuts across strongly folded rocks, and the formations above the unconformity also are deformed, there must have been two distinct epochs of orogeny. The several pulses of movement that affected the Alpine zone are differentiated by evidence of this kind.

Mountain making by any process is dated according to the same general principles. The latest episode of uplift in fault mountains cannot be older than the youngest rocks affected by the faulting.

The Ultimate Cause of Mountain Making

When it was believed that the Earth consisted of a relatively thin crust resting on a highly-heated liquid interior, it seemed easy to explain the origin of mountains by assuming that there was a regular contraction of the Earth's mass from loss of heat, and, as this contraction was greater in the heated interior than in the cold outer crust, the latter was folded up as it gradually sank upon the shrinking core, very much as the skin wrinkles upon a drying and contracting apple. This view can no longer be held, because the Earth is known to behave as a solid, rigid body; the interior cannot be wholly, or even chiefly fluid, in the ordinary meaning of the word.

Nevertheless a view commonly held to account for crustal deformation assumes contraction by loss of heat. This is a survival of the idea mentioned above, but changed to accord more nearly with later knowledge. It assumes that the Earth is solid and rigid but very hot within, and that progressive loss of heat causes slow shrinkage below a comparatively shallow depth. An English scientist has shown that this mechanism would account for a large amount of lateral thrusting in the outer shell throughout geologic history, provided we make certain reasonable assumptions as to the temperatures in the interior. Another form of the contraction hypothesis discards the idea of cooling, but assumes that the enormous pressures deep in the Earth cause matter to increase in density and decrease in volume. The net effect would be the same as if cooling occurred, as the outer shell, unchanged in volume, would collapse on the shrinking core, with consequent folding and thrusting. In either form the hypothesis encounters grave mechanical difficulties; among them, the necessity for explaining localization of deformation, during any one period, in only two or three widely separated geosynclines. It seems more probable that uniform contraction of the Earth would cause moderate failure of the crust in many zones, instead of violent deformation in a few.

It has been suggested that vertical adjustments in the outer crust to maintain isostatic equilibrium may be a sufficient cause of deformation. This suggestion does not find support in the evidences of enormous lateral pressure and displacement. In some mountain belts individual rock sheets have been thrust horizontally for distances of 25 miles or more; and every large folded geosyncline represents lateral movement through tens of miles. No force acting vertically in the crust could have a horizontal component so large. Moreover the fact of concentrated deformation in a few long, narrow belts is as difficult to explain by isostasy as by the contraction hypothesis.

Within the last few years some geologists have suggested that whole continents may shift horizontally through long distances. It is claimed, for example, that Africa moved northward against the old Mediterranean geosyncline and crushed it to form the Alps and neighboring mountains; and that the great folded chains of Asia were caused by southward shifting of that continent. It is urged that no other explanation will suffice in view of the stupendous shortening recorded by mountain folds and thrusts. But even if we should admit the moving of continents, the fundamental problem of orogeny would remain unsolved so long as the ultimate forces and conditions to cause such movement are wholly unknown.

It must be admitted, therefore, that the cause of compressive deformation in the Earth's crust is one of the great mysteries of science and can

be discussed only in a speculative way. The lack of definite knowledge on the subject is emphasized by the great diversity and contradictory character of attempted explanations. It is a fascinating problem, but lengthy discussion of its various aspects has no place in this volume. The facts and relationships of mountain structure present a large field of study in themselves, aside from the problem of ultimate forces.

Fault mountains that are due chiefly to thrust faulting present problems similar to those encountered in mountains with folded structure, as the fundamental cause is enormous lateral compression. Some steep faults that bound mountain blocks appear to be related to irregular shifting of magmas at depth, during the formation of intrusive or extrusive igneous bodies; and others are explained by vertical movements to restore or maintain isostatic balance with shift of load. Still others are formed in or near belts of severe folding, by irregular local twisting of the crust during the period of compression; and as fault mountains made in this way are merely incidental effects of the larger deformation, they become a part of the greater problem. Many dome mountains certainly owe their formation to igneous intrusion, and therefore the problem of their ultimate cause is closely linked with the problems of igneous activity. Some of the larger dome mountains appear to be closely related in origin to great folded units, and therefore involve similar uncertainties as to origin.

Summary of Mountain History

Mountains may be classified generally according to types of structure, which reflect different processes and forces that acted on the mountain zone. The visible structure may be due to simple dislocation and tilting of blocks, to simple doming of rocks, to folding with or without faulting, to large thrust movements, or to various combinations of these diverse processes. Connected with any type of movement there are commonly injections and extrusions of igneous material, which complicate the final structure of the mountain mass. Nearly all mountain-building movements take place in a series of pulses or phases distributed over a very long time. This is especially true of the chains that have complex structure. The cause of the great lateral pressure to which the belts of folding and thrusting bear eloquent testimony is an unsolved problem. We only know that long belts of weakness in the crust indicate their presence first by continued subsidence as geosynclines, and finally by yielding to forces that make the structures characteristic of the great mountains. Actual uplift to mountain heights, however, commonly follows the period of folding and appears to be largely independent of it.

Erosion is active in mountains throughout their history, and leads to progressive evolution of their varied forms. This aspect of mountain history is discussed systematically in the following chapter.

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Not up-to-date on all points, but has good descriptions, well written.

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CHAPTER XVII

LAND FORMS

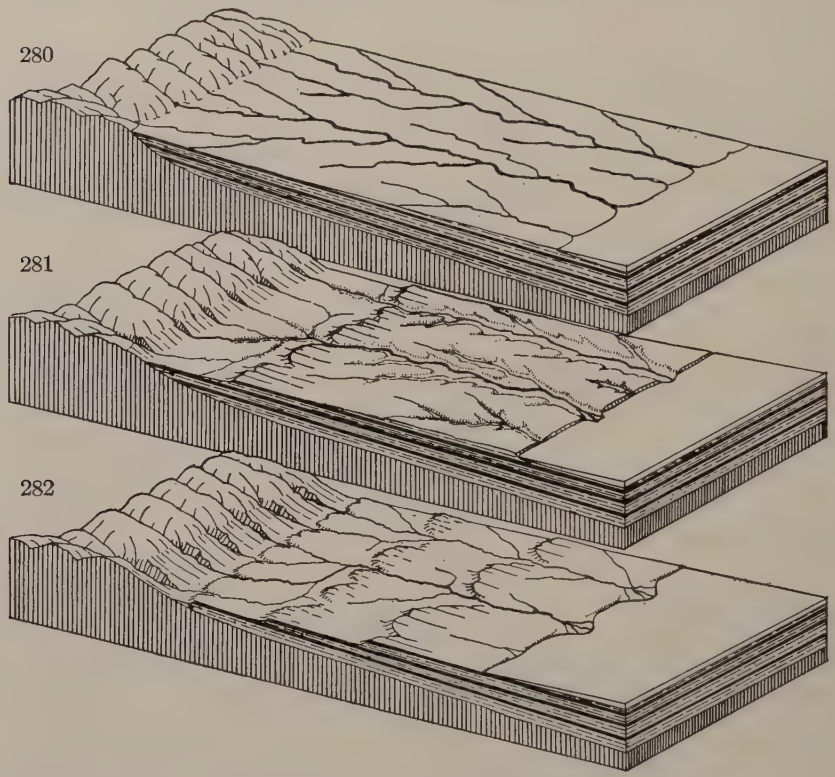
THE CYCLE OF STREAM EROSION ON TILTED STRATA

Extended Streams. — It has already been shown (Fig. 163) that when the sea encroaches upon a subsiding land mass, stratified deposits of different kinds are laid down with gentle seaward inclination. During the maximum submergence the shore deposits rest against the partly submerged *oldland.* When the sea has retreated from the area, the result is an uplifted coastal plain built of gently-dipping sedimentary rocks. The plain is not unbroken, however, because the streams of the oldland must have extended themselves mouthward as the sea began to retreat, in order to keep pace with the retreating shoreline, cutting at the same time an ever lengthening trench in the newly exposed coastal-plain rocks. Such streams are known as *extended streams.* They are of course consequent upon the uplifted surface as well (page 73); hence they are called *extended consequents* (Fig. 280).

Subsequent Streams; Cuestas. — As rain falls on the exposed coastal plain, tributaries develop, working headward from the trenches of the extended consequents. They develop most readily along zones of weakness in the underlying rocks. Such a zone is found in the contact of the recent shore deposits upon the much older, harder rocks of the oldland. Along this former land margin, then, the tributaries rapidly develop, working headward in a direction at right angles to that of the main streams. Such tributaries are called *subsequents;* all streams that work headward in this way along weak rock belts are so called. The parallelism of the weak belts in this case causes the subsequents to enter the main streams at right angles. The rectilinear drainage pattern thus developed is called a *trellis* pattern. (Compare the *dendritic* pattern developed in homogeneous rocks, page 76.) These subsequent streams cut the seaward, soft-rock sides of their valleys much more rapidly than the oldland sides where the rock is hard. Hence their channels constantly tend to shift seaward as by erosion they etch away the edge of the uppermost coastal-plain stratum. The product of this etching is a strongly asymmetrical low ridge, with a steep escarpment facing landward and a very gentle slope facing seaward formed by the surface of the

gently dipping coastal-plain deposits (Fig. 281). Such a ridge is known as a *cuesta* (the Spanish word for this type of land form).

Belted Coastal Plain. — Wherever the extended consequents discover a resistant layer overlying a weaker layer in the sedimentary series, they



Figs. 280–282. — Development of relief on a newly uplifted coastal plain.

Fig. 280. — Newly uplifted coastal plain composed of gently dipping beds of unequal resistance and showing progressive overlap, and drained by extended consequent streams.

Fig. 281. — Etching out of a cuesta from the topmost resistant bed by the development of subsequent streams.

Fig. 282. — Evolution into a belted coastal plain by the development of additional cuestas as lower strata are exposed.

send out subsequents which cut into the underlying soft rock, undermine the hard cap, and etch it slowly down its dip toward the sea. In this way a whole succession of cuestas may be developed, separated by the troughs of subsequent streams. Coastal plains bearing two or more cuestas are called *belted coastal plains* (Fig. 282).

The escarpment of a cuesta increases in height until the main consequents have become graded. At this stage downcutting by the tribu-

taries has virtually ceased, and since they can thus cut only laterally, the cuesta escarpments are gradually lowered by being etched seaward, until they disappear. The result is a peneplain.

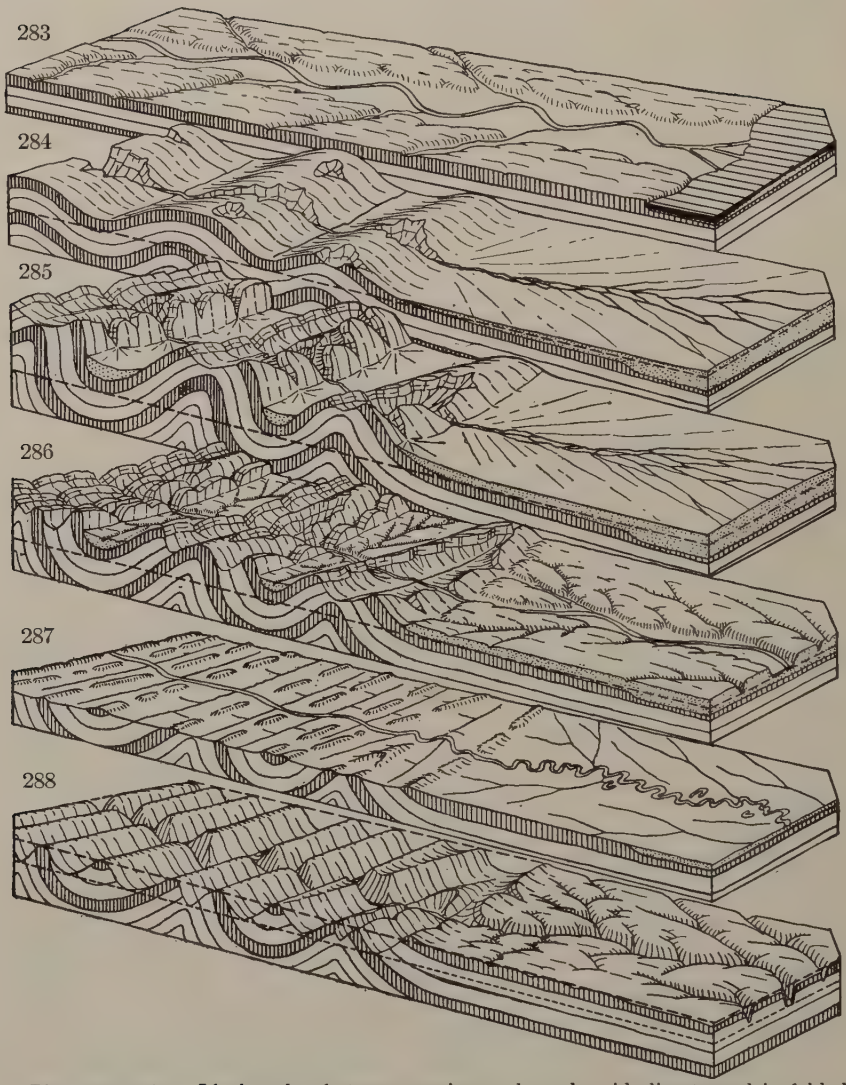
The coastal plain of New Jersey is belted, one of the belts being formed by the Navesink Highlands. In southern England, the North Downs and South Downs form two well developed series of cuestas facing each other across the Weald, a broad gentle anticline. Much of the country between Paris and the eastern frontier of France consists of a series of cuestas with east-facing escarpments. This arrangement was of the highest importance to the Allied armies on the western front during the Great War, since it presented a series of difficult natural obstacles to the German advance.

THE CYCLE OF STREAM EROSION IN FOLDED STRATA

Under humid climatic conditions, the denudation of an area of folded strata follows exactly the same laws as in the case of horizontal or homogeneous rocks outlined in Chapter IV. Local complications, however, are introduced because of the unequal resistance to erosion of alternating hard and soft beds. A typical case is outlined below which in many respects is similar to the early erosional history of the folded Appalachian ranges. Many variations of course are possible, but the following account is representative of the cycle of erosion and deposition under these conditions. In connection with this account, reference should constantly be made to Figs. 283–288.

Initial Stage. — Let the scene be set with a land mass in old age (essentially a peneplain) developed on a series of horizontal sedimentary rocks, alternately resistant and weak, of which two upper resistant beds with included weak beds are to be later affected by erosion (Fig. 283). The area is drained eastward by one main stream which reaches the sea over a well-developed flood plain and delta. The tributaries, because the rock is horizontal, form a typical dendritic pattern.

Stage of Early Youth; Antecedent Streams and Structural Valleys. — As folding begins, this section of the crust is thrown into three gentle anticlines (Fig. 284), of which the first plunges northward. The increased gradient attendant upon the folding brings the sluggish main stream to life, greatly increases its erosive power, and allows it to trench downward as rapidly as the anticlinal arches are uplifted. A stream which is strong enough thus to maintain its former course in spite of gradual uplifts across its path is said to be *antecedent*. Columbia River is antecedent to the great uplifted arch of the Cascade Range in Oregon and Washington. The tributaries, however, smaller and weaker than their



Figs. 283–288. — Ideal cycle of stream erosion under a humid climate and in folded strata. Each block is oriented so that its right end faces east and its long upper edge faces north.

Fig. 283. — Initial stage, showing a land mass in old age.

Fig. 284. — Stage of early youth, showing the beginning of folding and the development of an antecedent main with consequent tributaries in a trellis pattern.

Fig. 285. — Stage of later youth, showing valley fill, extensive breaching of the anticlines, and the development of subsequent valleys bordered by hogbacks.

Fig. 286. — Stage of maturity, showing dissection of the earlier valley fill, decrease in relief, and the development of rounded ridges within the anticlines.

Fig. 287. — Stage of old age, showing the weak rocks reduced nearly to baselevel, leaving only monadnocks of resistant rock for the streams to cut. Essentially a peneplain.

Fig. 288. — Stage of maturity in a second cycle instituted by regional uplift and rejuvenation. (See page 414.)

main, fare less well; they are blocked off by the folds and are extinguished. But new drainage lines are provided by the synclinal troughs. Concentration of drainage takes place in each of these, and pairs of tributaries are thus developed at right angles to the main stream, as consequents upon the folded surface. In this way the dendritic stream pattern is extinguished and a right-angled, or *trellis* pattern, takes its place. The synclinal tributary valleys are termed *structural valleys* because their position is determined entirely by rock structure. At the

Fig. 289. — Lehigh Water Gap as seen from Slatington, Pa. (Barrell.)

same time, rapid run-off down the slopes of the anticlinal limbs forms consequent tributaries of the second order, which rapidly excavate deep gorges. Directly these tributaries cut through the surface bed of resistant rock, and penetrate the weak bed below, they begin to cut the latter, undermining the former, and the valleys are quickly widened. This process is characteristic of the early stages of the cycle, in that the weak rocks are attacked as widely as possible, while the hard rocks, escaping direct attack, fall by undermining. The main stream crosses the anticlinal arches through *water gaps* of its own cutting (Fig. 289), and the greatly increased volume of débris in transport is deposited in part upon the uplifted coastal plain. That the stream system is confronted by a long period of cutting is indicated by the volume of rock above the new baselevel (indicated by a dashed line, Fig. 284).

Stage of Later Youth; Hogbacks and Canoe-shaped Valleys. — In this case the folding and general uplift is depicted as having now reached its climax (Fig. 285). The crests have thereby attained their maximum

elevation above baselevel (dashed line, Fig. 285). But because much erosion has taken place since the uplift began, it is clear that the actual mountains can never be as high as might be inferred at a later time by projecting upward the stumps of the anticlinal limbs. Steepening of the anticlinal slopes (compare Fig. 285 with Fig. 284) has resulted in such greatly increased erosion by the tributaries of the second order that the streams in the synclinal valleys, whose gradients have not been steepened much since folding began, are not able to carry away the waste contributed to them. The synclinal valleys are therefore partly silted up with

Fig. 290. — A hogback near Gallup, New Mexico. Buggy near left base gives scale. (U. S. Geol. Surv.)

waste in the form of coalescent alluvial fans. The ridges formed by the outcropping hard bed resemble steep cuestas. Such ridges of steeply-dipping rocks are termed *hogbacks*. They are common in the Colorado Rockies, the Bighorn Mountains, the Black Hills, and many other anticlinal mountain masses (Fig. 290). Whereas the anticline in the center (Fig. 285) has been breached along its entire crest, the anticline to the east is only partially breached, and that where its axis is highest. Since this is a plunging fold, its axis descends to the north, and hence the tributaries of the second order have lower gradients. This being the case, their cutting power is far less than that of their steeper neighbors, and the resistant bed which caps the anticline is therefore much less readily cut

through. This brings about the excavation of a *canoe-shaped valley*, the prow of the canoe pointing down the plunge of the anticlinal axis whereas its sides are formed by the anticlinal limbs. The canoe gradually widens as the retaining escarpments retreat. The whole gaunt structural skeleton is now dissected out in greatest relief, and the stream system has much cutting to accomplish before it can remove all of the rock above baselevel and carry it into the sea.

Stage of Maturity. — The weak strata continue to receive the most vigorous direct attack by the streams, and the resistant strata fall block by block as they are sapped and undercut. Both the canoe-shaped valleys and the others cut in the anticlines have been widened thereby (Fig. 286). Erosion of the soft strata in the middle anticline has now completely laid bare the rounded back of the lower resistant bed, which has already been breached to form an inner canoe-shaped valley. Inspection of Figs. 285 and 286 shows the appearance and development of other such rounded ridges at the hearts of the anticlines.

The whole land surface by this time has been reduced appreciably toward baselevel (dashed line, Fig. 286; compare Fig. 285). In fact, the heights have been lowered sufficiently to decrease greatly the supply of waste to the tributaries. The latter, freed from a part of their load of sediment in transport, are able again to devote a part of their energy to downcutting, and they begin to dissect their previous deposits in the synclinal valleys and on the coastal plain, cutting them into pronounced terraces. The main stream has now come so close to baselevel that it can cut down but little further. Hence it is easily diverted, and it begins to meander widely, broadening its valley by lateral planation. In this it is slowly followed by the tributaries in the synclines.

Stage of Old Age; Beveling of Resistant and Weak Rocks Alike. — The increasing predominance of lateral planation during late maturity and old age has resulted in the virtual cutting away of the remaining highlands. The reason is simply that all of the soft rock has already been worn down essentially to baselevel, leaving nothing but hard rock for the streams to cut (Fig. 287; note that the dashed line has disappeared because it practically coincides with the old age surface). Because of the great resistance of the only erodable rock and the low gradients, the process of cutting is almost infinitely slow. Figure 287 depicts a peneplain carrying elongate monadnocks formed by outcropping edges of the resistant beds, together with one rounded ridge in the plunging anticline. The structure is essentially beveled, hard and soft alike. The cycle is virtually ended; from now on the streams can work only to destroy the monadnocks, with scarcely appreciable results.

EFFECTS OF CHANGE OF LEVEL ON A LAND MASS

1. **Downward Movement; Drowning.** — Downward warping of the crust beneath any part of a stream's course causes a local decrease in gradient which may be followed by extensive deposition. The effects of downward movements are most plainly visible near the mouths of streams, which, if their channels are depressed below baselevel, are said

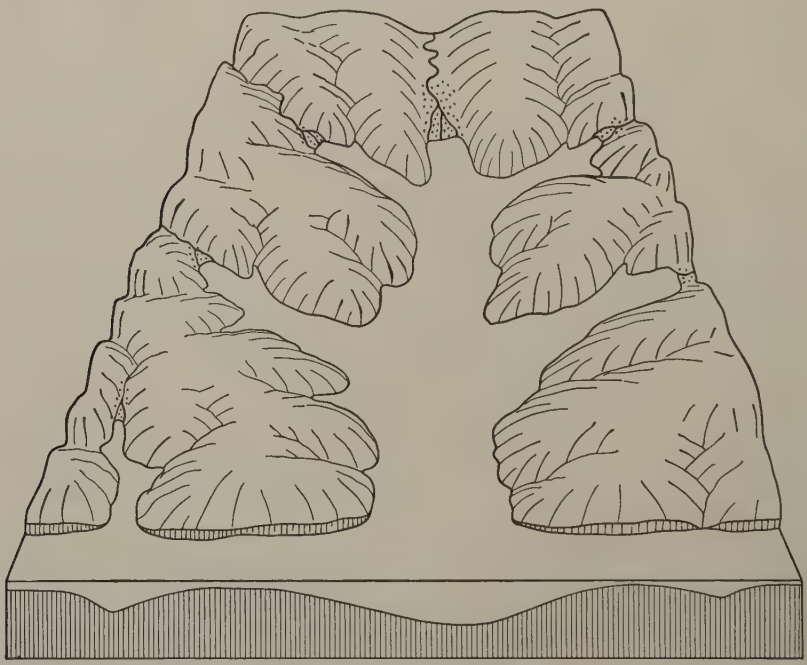


Fig. 291. — A river drowned in its lower reaches, forming a branching estuary. The estuary is slowly being destroyed by encroaching deltas.

to be *drowned.* Most of the larger rivers of the Atlantic Coast of the United States are drowned at their mouths. The lower Hudson occupies a deep narrow drowned valley cut in resistant rocks, and Chesapeake Bay represents the drowned lower course of the Susquehanna and the Potomac combined. The bay is broad and shallow because cut in weak rocks. The mouths of streams in southwestern Britain and northwestern France likewise are drowned over wide areas. Typical estuaries formed by the submergence of a valley are illustrated in Fig. 291.

2. **Uplift; Rejuvenated Streams and Entrenched Meanders.** — Upwarping of the crust is likely to increase the cutting power of the streams affected by increasing their gradients. Streams thus affected are said to

be *rejuvenated.* The effects, however, upon the aspect of a valley will be far more noticeable in the case of an old or mature stream rejuvenated, than in the case of one rejuvenated when still in its youth (Fig. 292). This follows from the fact that renewed cutting normally trenches a new V-shaped valley in the floor of the old one, and that if the latter is already

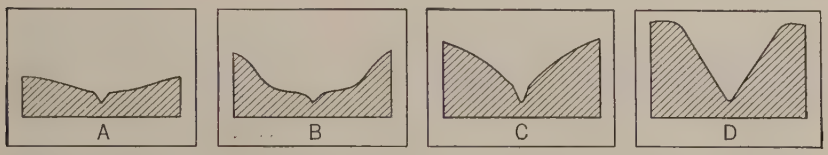


Fig. 292. — Cross profiles of valleys showing that the effect of uplift and rejuvenation depends on the stage at which it occurs.
A — Old stream rejuvenated: effect very strongly marked.
B — Mature stream rejuvenated: effect strongly marked.
C — Young stream rejuvenated: effect barely perceptible in the spurs along the valley sides.
D — Very young stream rejuvenated: effect not registered. (Drawn by E. J. Lees.)

V-shaped, its appearance in cross profile is scarcely altered. A keen eye can sometimes detect the evidence of uplift in a valley still youthful (Figs. 292 *C*, 293).

If a meandering stream is affected by uplift so rapid that the rate of lateral cutting by the stream cannot keep pace with it, the stream will be

Fig. 293. — East end of Boulder Canyon, southern Nevada. Note abrupt changes in slope of valley walls, suggesting at least two rejuvenations. (U. S. Geol. Surv.)

rejuvenated in the exact meandering course it followed immediately prior to the uplift. The meander loops thereby come to occupy gorges separated by steep-walled projecting spurs (Fig. 294). These loops are called *entrenched* meanders, and are a positive indication that uplift has

taken place in the region where they occur. Furthermore, the depth of the gorges is an approximate gauge of the amount of uplift.

The meanders cannot remain deeply and narrowly entrenched for long. The process of sweep is operative here as in the case of normal meanders (page 56). Lateral cutting is greatest against those sides of

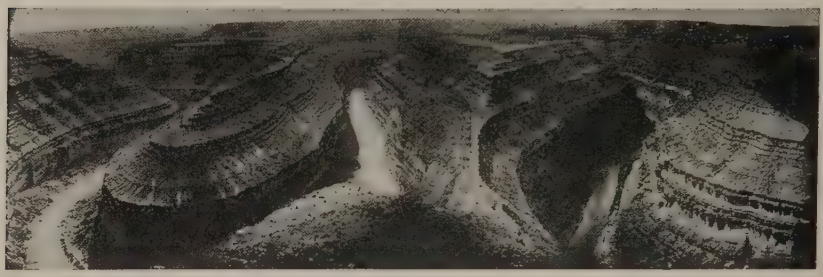


Fig. 294. — Entrenched meandering valley of the San Juan River, 30 miles below Bluff, Utah; canyon 1200 feet deep. (Vinson.)

the curves that face up against the general course of the river, since upon them the force of the current must come with attendant corrasion. These sides of the spurs interlocking between the meanders therefore

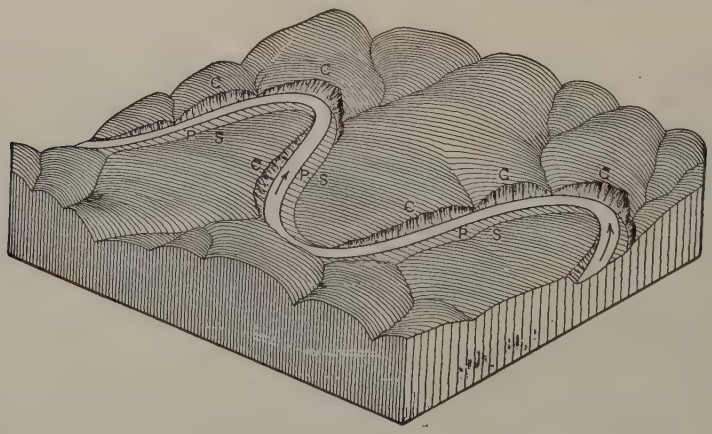


Fig. 295. — An entrenched meandering valley; arrows point downstream. *C*, *C*, *C*, undercut slopes of valley; *S*, *S*, *S*, slip-off slopes; *P*, *P*, *P*, flood-plain scrolls. (After Davis.)

tend to be undercut and to present steep and even cliff-like faces to tne course of the river (*C*, *C*, *C*, Fig. 295). The down-valley sides of the spurs, *S*, *S*, *S*, on the contrary are apt to descend to the river with more or less gentle and gradual slopes, commonly covered with sand or gravel deposited by the stream. The reason for the difference is that since the

river cuts laterally against the faces *C*, *C*, and also vertically downward, these sides of the spurs are being eaten into and consumed, whereas it tends to move away from the sides *S*, *S*, and with slack current to leave deposits on them. Since the stream tends to slide away from them without eroding, they are called *slip-off* slopes, as distinguished from the *undercut* slopes *C*, *C*. As one looks down the course of such a valley he sees only the steep and wooded or cliff-like faces of the undercut spurs, which give it a stern and rugged aspect; when he looks up the valley the cultivated fields of the gentle slip-off slopes confront him.

From what has been said above it will be seen that an entrenched winding river tends to become more circuitous in its course, and that the whole system of meanders moves down the valley. Further, when downcutting decreases and the stream becomes graded, it begins to build narrow strips of land or scrolls, *P*, *P*, *P* (Fig. 295), along the insides of the curves *S*, *S*, *S*. Material is gradually and steadily added to these scrolls, and they grow in size. As the valley widens by lateral planation they coalesce, and a continuous alluvial flat is established.

It is evident that as the meanders enlarge and change their curves they are likely to form short cuts, as shown in Fig. 26. In an entrenched meander it may happen that the neck of land connecting the spur end is actually undercut at the narrowest place, leaving the former spur remnant as an island joined to the mainland by a *natural bridge*, under which the stream runs in the short cut it has made.

The rivers of northeastern France and Belgium, such as the Meuse and the Moselle, furnish typical examples of entrenched meanders. The headwaters of the Susquehanna in Pennsylvania, the Kentucky, and many others, are equally good examples.

The completion of the ideal cycle outlined on pages 409–413 and depicted in Figs. 283 to 287 is followed in Fig. 288 by uplift and rejuvenation. In other words, the former baselevel (upper dashed line) is abandoned in favor of the new baselevel (lower dashed line). This permits the stream system to renew the work of excavation which it had completed under the old conditions. The weak rocks are again etched away from the resistant rocks, which are left standing as prominent ridges. The tops of these ridges are beveled flat, and they are worn down so slowly that during all of the early part of this new (second) cycle they preserve the level of the former peneplain from which the intervening weak rocks have been cut away. These ridges are said to be *peneplain remnants*, and can be formed only by uplift and the beginning of a second cycle following the completion of an earlier one. In an analogous manner, the highest ridges of the folded Appalachian ranges are remnants of a former peneplain. Their flat beveled tops betray the fact that they once stood at or close to

baselevel and that they were later etched into relief following a great uplift.

In Fig. 288 the weak rocks in the two synclines are temporarily preserved as ridges because protected by narrow caps of resistant rock. The latter however must disappear in time by being undermined, and the synclinal ridges will then be quickly destroyed. The completion of the second cycle will see a baseleveled surface (developed in the plane of the lower dashed line) much like that of Fig. 287. There will be two chief differences, however: (1) Monadnocks will be fewer because of the complete removal of the upper resistant layer from the two synclines. (2) The remaining monadnocks will be farther apart because the hard layers will be intersected *farther down their limbs* by the plane of the new baselevel. Thus the hills, strong bastions in appearance, are in reality slowly moving down the dip of the resistant strata of which they are composed.

Rock Terraces Caused by Successive Uplifts. — Rock terraces caused by differential erosion of horizontal strata are discussed in Chapter IV. Terraces are also not infrequently cut in tilted or completely folded strata, beveling them regardless of their unequal resistance (Fig. 296). Since hard and soft beds alike can be brought down to a common plane

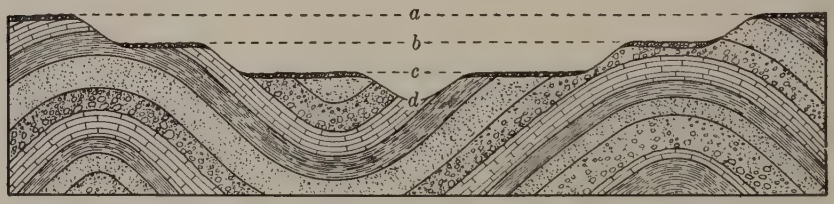


Fig. 296. — Rock terraces caused by successive uplifts. (After Wright. Va. Geol. Survey.) Terraces *a*, *b*, and *c*, are remnants of old graded valley flats, elevated by three successive uplifts, the last of which is causing the excavation of the trench *d*. Width of cross section is of the order of magnitude of $\frac{1}{2}$ mile. (Drawn by E. J. Lees.)

only near baselevel, it follows that each pair of rock terraces of this type must have been formed by a stream in the latter half of its cycle, and that for each pair of similar terraces there must have been a corresponding uplift of the land followed by renewed downcutting toward baselevel and then widening of the valley by lateral planation. Thus in Fig. 296, a nearly baseleveled surface *a* must have been uplifted and dissected, followed by valley widening near a new baselevel *b*. A second uplift resulted in the cutting of the plane *c* which after a third uplift was dissected into terraces by downcutting of the trench *d*. Obviously peneplanation took place in no case; if it had, the higher terraces would have been destroyed. Each terrace is veneered with a thin layer of

alluvium, the remains of an alluvial flat built up by continuous deposition on the inside curves of meanders.

Adjustment of Streams to Structure. — A close study of Figs. 283–286 shows the remarkable fact that although the initial tributaries develop as consequents in the synclines, they steadily decrease in importance until in old age (Fig. 286) they are almost completely extinguished. The

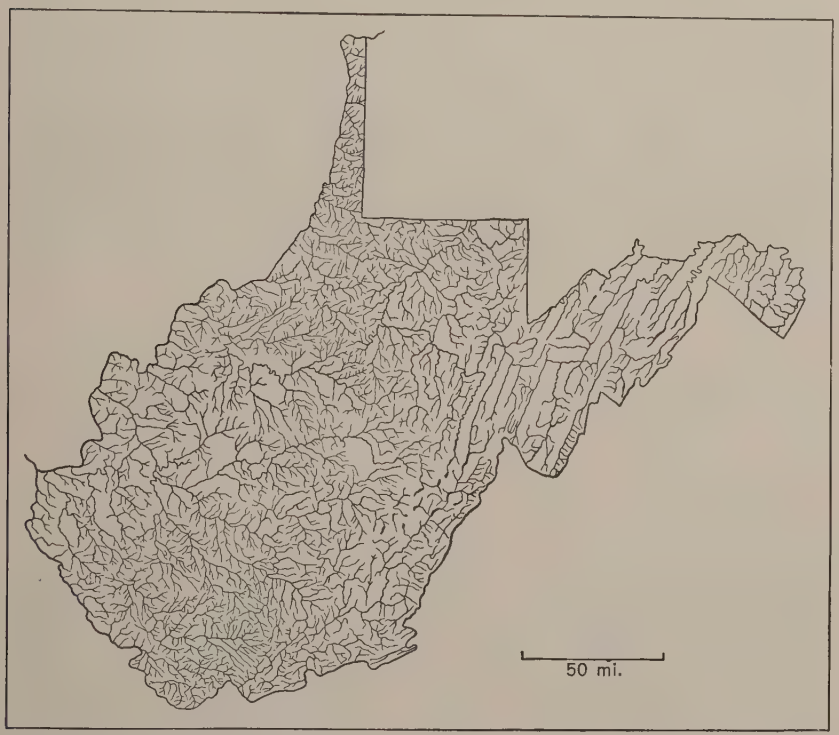


Fig. 297. — Outline map showing stream patterns in West Virginia. In the central and western regions the rocks are nearly horizontal and the pattern is dendritic. In the east the rocks are strongly folded along NE-SW axes and the resulting pattern is trellised. Note how thoroughly the region is drained (mature dissection under a humid climate), and compare Fig. 62, Chap. IV. (Drawn by D. Gallagher.)

further fact appears that as the resistant caps of the anticlines are breached, subsequent streams (page 407) are developed in the weak rocks below, parallel to the synclinal consequents. As the cycle progresses these subsequents gain in importance, so that before the end they have become the most important tributaries. This growth of the subsequents at the expense of the consequents is called *adjustment to structure* and is explained by the simple fact that the former can cut more rapidly in thin weak rock beds than the latter are able to cut in the

resistant rocks through which they flow. From this the universal principle may be set up that *stream systems are constantly forced to adjust their courses so as to flow as much as possible on weak rock and as little as possible on resistant rock.* Until this has been achieved within the limits of existing conditions, the streams are not completely adjusted. Drainage developed in folded strata rarely reaches an adjusted condition before late maturity or old age. The more or less complete adjustment of the larger streams in the folded Appalachians is illustrated in Fig. 297. From the trend of the trellised stream pattern the strike of the folds can be readily inferred.

Superimposed Streams. — Certain streams are so completely out of adjustment and pursue courses so wholly regardless of structure that some special set of conditions must have brought about this inharmonious relationship. Suppose the seaward portion of a peneplain developed on

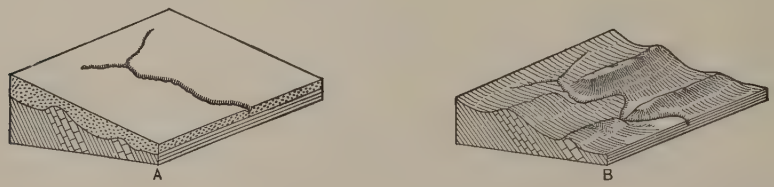


Fig. 298. — Development of a superimposed river. *A*, course determined by initial slope on a layer of sand and gravel; *B*, latter removed by erosion and river pursuing its course without regard to underlying rock structure.

folded strata of unequal resistance to be depressed slightly. During the resultant submergence, marine deposits are laid down almost horizontally upon the drowned erosion surface, forming an angular unconformity (page 332) with progressive overlap (page 234). When now the monotonous submarine plain is uplifted, consequent or extended streams develop on the gently sloping surface, exhibiting a dendritic drainage pattern (Fig. 298, *A*). In time the streams cut down through the unconformity into the different material and structure below. When the overlying sediments have been stripped away, the streams are found to be following unadjusted courses *inherited from the overlying unconformable strata* (Fig. 298, *B*). Such streams are said to be *superimposed.* Streams may be thus "let down from above" from glacial, volcanic, eolian, and alluvial deposits as well as from marine deposits. In any case, gradual adjustment will take place as the cycle progresses.

From the foregoing, it must be true that any stream which cuts across the structure of folded rocks, and which therefore cuts water gaps across tilted resistant strata, *must be either antecedent or superimposed.* Neither consequents nor subsequents could occupy such positions without first

becoming antecedent or superimposed. If remnants of overlying unconformable strata still cap the ridges through which the gaps are cut, the case for superimposition is fairly clear. Evidence of antecedent history is more obscure.

Wind Gaps; Capture by Subsequent Streams. — A water gap (Fig. 289) abandoned by its through-flowing stream is called a *wind gap* (Fig. 299).

Fig. 299. — Wind Gap in Kittatinny Mountain at Pen Argyl, 11 miles north of Easton, Pa. (Barrell.)

Most of the wind gaps which have been studied are the result of capture by subsequent streams. This process is best illustrated by the case of Snickers Gap, a prominent notch in the Blue Ridge of Virginia about

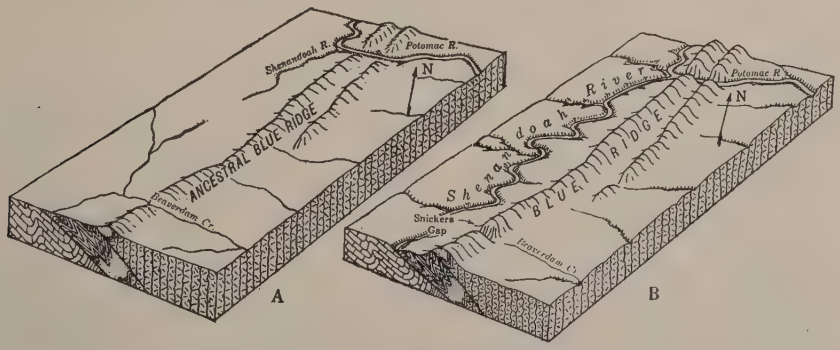


Fig. 300. — *A*, former drainage across the Blue Ridge in northern Virginia. *B*, present drainage resulting from the beheading of Beaverdam Creek by the Shenandoah.

fifteen miles south of Harpers Ferry. The floor of this notch hangs 700 feet above Shenandoah River at the west base of the ridge. A small stream called Beaverdam Creek heads near Snickersville at the east base, and flows eastward away from the ridge. The notch is a large one, and must have been cut by a large stream. Its reconstructed history is as follows: During an earlier cycle than the one now in progress, when the land around the base of the Blue Ridge stood as high as the notch of

Snickers Gap, this notch was occupied by a transverse stream, the ancestral Beaverdam Creek (Fig. 300, *A*). But the much larger transverse ancestral Potomac, which crossed the ridge 15 miles farther north, was able to cut its gap downward into the hard ridge-forming rocks much more rapidly. The Shenandoah, tributary to the Potomac through an easily eroded limestone area west of the ridge, kept pace with its master stream, and with the resulting favorable gradient, worked headward toward the south, cut through the divide which separated it from Beaverdam Creek, and "beheaded" the latter by diverting its upper waters toward the Potomac (Fig. 300, *B*). The gap was forthwith abandoned, and the beheaded and weakened creek found itself confined to the area east of the Blue Ridge. As the Shenandoah continued to cut downward, the ridge was etched out in greater relief and the abandoned gap was left high and dry.

Some wind gaps may have been formed by relatively weak streams which, affected by upwarps athwart their courses, were able to maintain their courses only for a time, having been later forced to abandon them and to flow along the strike.

EROSION IN RELATION TO MOUNTAIN MAKING

It has already been pointed out (Chapter XVI) that the work of erosion goes forward hand in hand with the uplift of great crustal masses, etching out forms of mountainous size. The work of erosion, then, is of the greatest importance in a full consideration of mountains. It begins with the first rising of the masses, proceeds while the orogenic forces are at work, and continues long after they have come to rest. As its results become especially marked in this last stage, it must be considered the chief agent in mountain development during the later phases of mountain history.

Earlier Stages of Erosion. — So long as the compressive orogenic forces are at work, a mountain range grows in so far as its structure is concerned. Whether it actually rises in height or not depends on the adjustment between (1) vertical uplift which tends to make it rise, and (2) the work of erosion which tends to cut it down. Always during the formative period this struggle goes on, and the height of the range at any time is a function of these two forces. When the orogenic movements cease, then denudation has full sway and, ultimately, with the lapse of time and provided no renewal takes place, the range must be cut down, baseleveled, and extinguished by the relentless agents of erosion. In this process various stages are to be distinguished. When the range is at its maximum elevation the erosive agencies are most

severe; to the work of running water on steep slopes is added very commonly the effect of frost, snow, and ice.

It may happen also that at this time the rock material exposed to erosion consists of the later beds laid down in the geosyncline, which have suffered less metamorphism than the deeper, older ones, and are thus less resistant to erosive attack. If igneous extrusions have contributed to swell the volume of the range, it will also be the more easily eroded tuffs and lavas that are first exposed. Hence, in general, the outer material is more easily cut away, and the inner core progressively exposes to erosion more and more resistant rock. Thus, in the early history of a range not only is the severity of attack of eroding forces likely to be increased by great height, but they may find less resistant material to work upon. At first the upraised masses begin to be trenched by the valleys of the initial consequent streams. The drainage lines thus appear upon original slopes and continue to cut downward and to work backward into the range. As they do so they begin to be conditioned more and more by the structure and nature of the underlying rocks.

Fig. 301. — Longitudinal profile of a mountain range in early maturity.

The mountain masses are profoundly graved and acquire rugged peaks and towering rock pinnacles, alternating with deeply scored valleys. The strongly notched outlines of such ranges present a saw-toothed appearance, which has led to their being called by the Spanish name of *sierra* (saw) (Fig. 301). The topographic development of a range thus proceeds from youth into early maturity, and as erosion continues and the valleys widen, the declivities lessen, angularities of form tend to disappear, and the mass becomes more and more mature. The topographic forms of the peaks and ridges and of the intervening valleys must depend largely on the nature and structure of the rock masses presented to erosion.

The Jura Mountains of Switzerland (Fig. 274) present a type of somewhat youthful dissection; here the folds themselves are the dominant topographic features, which erosion as yet has been unable to modify greatly. Many of the ranges in the Alps, the Himalayas, the Caucasus and the Rockies are in mature stages of dissection, and their folded strata have been breached in the anticlines and largely etched away. The terms *youthful* and *mature* in this connection are merely relative, and do not refer to absolute time; actually one range may be much older than

another, and yet on account of its greater mass, difference in material, or difference in climate, be in an earlier stage of its life history.

Later Stages of Erosion. — If erosive processes continue their work of degrading a mountain mass, unhampered by further uplifts, the range gradually passes into a mature stage. The sharp peaks and asperities tend to disappear, the valleys to widen. The progress of the work goes more slowly as the slopes lessen, and as the resistant metamorphic and crystalline rocks of the inner core are reached. Thus a maturely dissected range presents rather smoothly rounded forms and outlines (Fig. 302), which contrast sharply with the angular features of the sierra type. As they wear down more and more and pass gradually into old age, we find these mountains in humid climates composed of massive rocks, of schists, gneisses, and granites, rather than of the limestones, sandstones, shales, and lavas of ranges in the earlier stages. In arid climates, on the other hand, limestones outlast the granitic rocks because

Fig. 302. — Characteristic forms and outlines of late-mature mountains.

they are far less vulnerable to the mechanical weathering that characterizes dry regions.

There appears to be no well-recognized term equivalent to *sierra* for mountains in these later stages; they are variously termed mature, subdued, or old mountains. Examples are to be seen in the mountain masses of New England and eastern Canada, such as the Green Mountains, the White Mountains, and the Laurentian Mountains of Quebec; in Europe the Black Forest region of Germany and the Highlands of Scotland and Norway are examples.

Final Stage: Peneplanation. — Ultimately, provided no new upwarping movements occur, the mountains will disappear and the region they occupied will be reduced nearly to baselevel. Since, however, the process of erosion goes on more and more slowly as the slopes lessen, it would evidently require an enormous lapse of time to bring down actually to baselevel a mountainous tract, and we have no proof that this has ever in fact occurred. But we know that some areas have been reduced to low, almost featureless country; in other words to peneplains. Such country may still be diversified by scattered monadnocks projecting above the general level which, on account of their more resistant composition, or possibly because of their position far from large streams, have not been reduced like their neighbors (Fig. 303).

Disregarding occasional monadnocks, we may say that when the

peneplain stage is reached the mountains have been obliterated; but we may yet be able to infer their former existence by the upturned and dislocated nature of the transversely eroded strata, by the widespread metamorphism of the rocks, by the slaty cleavage and faults which cut them, and by the presence of large granitic intrusive masses. We can not determine the former elevations, for, as LeConte has said, "we find only the bones of the extinct mountains"; but from these remains we may learn the trend and extent of the ancient ranges. So, from the attitude of the rocks of southern New England, which is now only a hilly country, we are led to infer that it was once a mountainous region.

Fig. 303. — Stone Mountain, De Kalb Co., Georgia. A monadnock composed of granite which rises above the surrounding plain of erosion. (Geol. Surv. of Georgia.)

Reëlevation; Complexity of Mountain History. — If mountains were forever extinguished by the peneplanation at the end of a cycle, their history would be comparatively simple. But though the surface of the land may be smoothed out by erosion, the structure below the surface remains. The tilted and folded strata of unequal hardness and the igneous intrusions injected into them are still there, having disappeared as relief features because of peneplanation, but needing only a second uplift of the land to bring them again into prominence. No new deformation of the rocks is required. A simple upwarping of the beveled surface, such as has occurred again and again throughout the Earth's history, accomplishes the result. The sluggish streams are rejuvenated; they begin to cut actively, and *they cut the weaker rocks most rapidly*, leaving the resistant masses once more projecting above the surface. The moun-

tains are thus etched again into relief, their new height depending entirely upon the amount and rate of the new uplift.

Here is a range "in its second cycle," etched by streams from an uplifted peneplain. How is the geologist to recognize the twofold nature of its history? The one indicator that is present in all mountains during the earlier stages of their second cycle is the close accordance of their summit levels, representing the surface of the former peneplain (Fig. 288). These summits will not endure throughout the second cycle, but they will be the last to be brought down toward the new baselevel because they are made of the most resistant rocks.

Many ranges, however, have passed through an even more complicated history. Not a few have suffered uplift and denudation repeatedly. If each period of denudation had resulted in a peneplain, it would be impossible to unravel the complete chain of events in such an intricate history; but wherever renewed uplift began to affect a mass well before the end of a cycle of erosion, some of the summits etched out during the earlier cycle were spared by the later, and thus old summits and broad rock terraces (page 418) were left as witnesses to what had happened.

History of the Appalachians. — Through close adherence to these principles, it has been possible to reconstruct the history of the Appala-

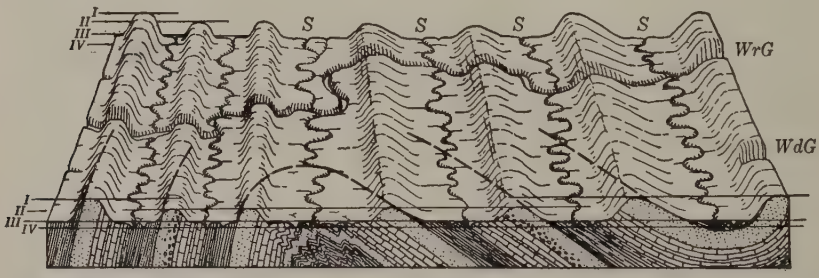


Fig. 304. — Ideal section across a part of the northern folded Appalachian ranges. The main stream is either superimposed or antecedent, trenching the hard ridges through seven water gaps (*WrG*). A former stream course, long since abandoned, is indicated by a wind gap (*WdG*). The present tributaries are subsequents (*SS*). Four successive baselevels (*I, II, III, IV*) are indicated by four accordant series of summit levels and valley floors, thus indicating four regional uplifts relative to sea level. Compare Fig. 288.

chians. As we see them today they consist chiefly of long parallel ridges formed by the outcrops of very resistant rocks such as sandstones and conglomerates in a strongly folded sedimentary series (Fig. 304). Most of the ridges, even though narrow, have remarkably level summits, broken only occasionally by water gaps (*WrG*) and wind gaps (*WdG*). Even a casual observer might notice that groups of these ridges reach a common level (*II*, Fig. 304), and if he examined them more closely he could see

that certain of them reached a notably uniform higher level (*I*, Fig. 304). And if our observer became really interested and gave some attention to the broad valleys (*III*, Fig. 304) between the ridges he would discover what appear to be old valley floors deeply dissected by small streams into a network of low hills. The small streams drain into larger meandering subsequents (SSS, Fig. 304), flowing down the valley axes, *and all the meanders are entrenched.* The subsequents in turn are tributary, in the northern and central Appalachians, to great streams which flow eastward indiscriminately across hard-rock ridges and soft-rock valleys. These are represented by the Delaware, the Susquehanna, the Potomac, the James, and the Roanoke.

With these facts in mind, how much can we reconstruct of the Appalachians' history? *First,* a period of strong deformation, as revealed by the great folds whose eroded stumps appear in the ridges. *Second,* erosion which must have begun during the first uplift, and have continued throughout a long cycle, resulting at length in a peneplain. This we can tell because the hard rocks in the folded series have been beveled down to a common level, as in the ridges *I* and in many others like them. Of course these rocks did not exist as ridges during the peneplain stage, but were merely parts of a low, gently undulating plain near sea level. *Third,* slow bodily uplift of the whole region, with rejuvenation, etching out of the resistant rocks into ridges again, and eventual reduction of the whole mass into a second peneplain at level *II.* The most resistant or most favored ridges (*I*) were not reduced, but remained as long narrow monadnocks. *Fourth,* a second slow upwarping, with the inevitable reëtching of hard ridges from the peneplain *II,* and the development of a new baseleveled surface *III* on the soft rocks first attacked. This new surface is scarcely a true peneplain because of the great quantity of hard rock (all the *I* and *II* ridges) still unreduced. The meandering streams were just beginning to destroy these ridges when the third uplift occurred and forced them again to cut straight downward. *Fifth,* a third uplift of the land, rejuvenating the streams and thus causing them to entrench their meanders and with the aid of their tributaries to dissect the old baseleveled valleys *III* into the network of hills now existing.

This, stripped of complicating minor movements and events, is the accepted history of the Appalachians; and if streams have behaved in the past as we see them behaving today, this history must be true. Authorities are not as yet agreed as to the exact time at which each peneplain was formed, but the whole sequence of events from the orogenic period down to the present may have required 200 million years.

FACTORS INFLUENCING DENUDATION

Climatic Control of Denudation. — The stages of the cycle of stream erosion under arid conditions have already been outlined (page 85). Complications of structure of course introduce changes in the progress of erosion and deposition. The effects, however, are less pronounced in an arid climate because of the shortness and rapid disappearance of the streams and because of the masking effect of the predominant alluvial fans. The important consideration here is that changes of climate bring about great changes in the landscape developed on any given set of structural conditions. Thus a change from a moist climate to a dry in a mountainous region causes the development of bolson basins; or an equally great change from a temperate climate to a cold may freeze the water into perennial ice and thus bring on glaciation with its resulting characteristic landscape.

Structural and Lithologic Control of Denudation. — The structure and lithology of a land mass are of vital importance to the landscape resulting from any given dynamic process such as stream erosion. The landscape developed in a mass of weak rocks elevated only slightly above the sea can never be more than one of rolling monotony even in maturity when relief is greatest. Rapid weathering and corrasion will provide an abundance of waste, the deposition of which, combined with lateral planation (much downcutting being impossible) results in the ready development of wide, open valleys. Similar streams will be able to act very differently, however, on a mass of resistant rocks elevated high above sea level. Steep gradients permit the development of scenic gorges whose sidewalls are accentuated in their steepness by the resistance of the hard rock to weathering and slope wash. Moreover, the slow inevitable changes wrought by erosion during the progress of a cycle in any land mass and under any climate, steadily alter the surface until the change has become profound. Thus it appears that the seemingly unending variety of landscape is in reality controlled by a few simple factors, and that slight variations in these factors produce the differences in scenery which lend enjoyment to travel and add to the richness of human existence.

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CHAPTER XVIII

ORE DEPOSITS

Man wrests from the Earth many materials of the mineral kingdom for his necessities of life and comfort. The search for them has given rise to romance and adventure; their discovery has resulted in the opening up and settlement of new countries; their ownership has resulted in national, political, and commercial supremacy or has caused strife and war. Their richness has often been the incentive for man's acquisitiveness. In the quest for these substances it is necessary to know about their distribution, occurrence, character, and origin, all of which is a part of the science of *economic geology*. Economic geology deals also with problems of investigation and other applications of geology to the uses of man, but we are concerned here merely with that part relating to mineral deposits.

Of the great variety of mineral substances won from the Earth, coal is the most valuable, followed by the metallic minerals, petroleum and natural gas, and the nonmetallic minerals such as salt and feldspar. But since coal and petroleum are treated in Part II of this book, and many of the nonmetallic minerals are considered briefly in the different chapters, we shall restrict ourselves to the important group of metallic mineral deposits or, as they are commonly called, *ore deposits*.

Ore deposits are geologic bodies that may be worked commercially for one or more metals. They are exceptional features, sparsely scattered in the rocks or on the surface; they constitute only an infinitesimal part of the Earth's crust, but they assume an importance far in excess of their relative volume because of the highly valuable materials they supply to natural wealth and industry. They have been concentrated in the rocks under peculiar and exceptional conditions that it will be our purpose to study. They cannot properly be considered apart from their geologic environment; consequently the information contained in the preceding chapters is vital to an understanding of them. A knowledge also of the origin and character of an ore deposit may aid in prophesying its size and depth beneath the outcrop.

MATERIALS OF ORE DEPOSITS

An ore deposit is of value for the metal or metals it contains, and these are usually locked up in one or more *ore minerals*. The latter in turn are

commonly admixed with *gangue minerals*, and the mixture, which constitutes the *ore* is inclosed in the *country rock*. The term ore is often loosely used to designate anything that is mined from the Earth; but in a technical sense it denotes that part of a geologic body from which the metal or metals it contains may be extracted profitably. Nonmetallic substances such as coal, salt, feldspar, or building stone, which are used practically in the form in which they are extracted from the Earth, are thus excluded. A lead ore deposit, for example, may be inclosed in limestone country rock; the lead is chemically combined with sulphur in the ore mineral galena, and the latter may be admixed with the gangue mineral quartz, to form the ore. The winning of the lead from the ore deposit involves first a knowledge of the occurrence, shape, continuity, content, origin, and other geologic features of the deposit. Thus an understanding of the geology of the deposit is usually prerequisite for intelligent mining operations. The study of ore deposits, it will be seen, is preliminary to the other steps in the winning of metals.

Ore Minerals. — An ore mineral is one that may be used to obtain one or more metals. Thus galena is an ore mineral because it is mined for its metallic lead; but feldspar, although containing as much as 15 per cent of aluminum, is not an ore mineral because it is not mined for its aluminum content. The ore minerals occur as *native metals* or as chemical combinations of the metals with other elements. Gold and platinum usually occur as the native metals, and silver and copper are often found in that state. Most of the common metals are chemically combined with sulphur, arsenic, carbon, oxygen, or silica.

Some ore minerals contain two or more metals, as, for example, chalcopyrite with its copper and iron, and individual metals may enter combinations to form several different ore minerals. In addition, several metals as, for example, silver, lead, and zinc, may occur in one deposit. Thus it is clear that the ore minerals of a deposit may represent a complex mixture of several metals, occurring in several different combinations, and with a single metal in more than one combination. In certain ore deposits, however, such as those of iron, this is not the case; only the one metal, iron, is obtained, and this occurs in just the one combination — iron oxide.

The metals of commerce are derived from many metallic combinations. Most of the world's gold has come from the native metal; consequently its removal from ores is a relatively simple process and offered no serious problem of extraction even to the ancients. Silver, on the other hand, is derived not only from the native metal but also from its combination with sulphur (sulphide). This is also true of the copper of commerce. Lead and zinc, however, are obtained chiefly from minerals containing

sulphur, although combinations with carbon and oxygen contribute an appreciable amount. The vast quantity of iron used in industry is obtained almost entirely from combinations with oxygen (oxides). The simple metallic combinations just enumerated have yielded pure metals readily to the art of extraction and have supplied the human race with metals for over 2000 years. It must not be overlooked, however, that other less important and more complex metallic combinations not mentioned above yield appreciable amounts of the common metals as well as many of the minor metals not considered in this chapter.

Some of the important ore minerals from which the common metals are extracted are listed below, and several of these are described briefly in Appendix A.

List of the Commoner Ore Materials

Metal	Ore Mineral	Composition	Percentage of Metal
Gold	Native gold	Gold	100
Silver	Native silver	Silver	100
	Argentite	Silver, sulphur	87
Copper	Native copper	Copper	100
	Chalcopyrite	Copper, iron, sulphur	34
	Bornite	Copper, iron, sulphur	55
	Chalcocite	Copper, sulphur	80
	Cuprite	Copper, oxygen	89
	Malachite	Copper, carbon, oxygen, water	57
	Azurite	Copper, carbon, oxygen, water	55
Lead	Galena	Lead, sulphur	86
	Cerussite	Lead, carbon, oxygen	77
	Anglesite	Lead, sulphur, oxygen	68
Zinc	Sphalerite	Zinc, sulphur	67
	Smithsonite	Zinc, carbon, oxygen	52
	Calamine	Zinc, silica	55
	Zincite	Zinc, oxygen	80
Iron	Magnetite	Iron, oxygen	72
	Hematite	Iron, oxygen	70
	Limonite	Iron, oxygen, water,	60
	Siderite	Iron, carbon, oxygen	48

Gangue Minerals. — Gangue minerals are the valueless minerals of the ore, and are usually earthy or nonmetallic in character. In common usage they are simply referred to as gangue — an old mining term. Thus in an ore deposit containing quartz and galena, quartz is the gangue. Several gangue minerals may be present in one deposit. Some of the common gangue minerals are quartz, calcite, dolomite, siderite, and limonite.

Gangue minerals are lacking in some deposits, and then the inclosing country rock is sometimes loosely referred to as gangue, so the term is somewhat flexible. Some of the gangue minerals considered worthless today may under improved metallurgical processes turn out to be valuable ore minerals of tomorrow.

The Ore. — It will be evident from the foregoing statements that ores vary greatly in mineral content and chemical composition. No two are ever exactly alike. There are simple ones — iron, for example, — which contain no gangue minerals and are composed solely of the ore mineral hematite. This is commonly thought to be the case with the ores of other metals, but it is far from correct. The ore minerals usually constitute but an insignificant part of the ore. If one were to take a trip through a profitable gold mine he might search the ore in vain to see a speck of gold; gangue minerals alone would meet the eye, since the weight of gold may form only a few ten-thousandths of one per cent of the ore. The proportion of ore to gangue minerals in deposits of copper, lead, and zinc, however, falls between the extremes of iron and gold, but the gangue usually predominates.

The abundance of metals in ores is also reflected in their relative prices, for a ton of iron may be purchased at about the same price as an ounce of gold. The higher the value of the metal the lower the grade of ore that can be mined for the same cost. Thus, metals such as gold or platinum make profitable ore with only a few tenths of an ounce of metal in each ton of ore. It is evident that the *tenor* or metallic content of ore that may be profitably mined depends largely upon the selling price of the different metals, the size and character of the deposits, and their accessibility. Thus economic as well as geologic factors determine what constitutes ore. The tenor obviously varies with deposits of different metals and with different deposits of the same metal. Man, of course, imposes no upper economic limit to the tenor of the ore; the richer the better. But the lower limit is vital, since sufficient metal must be extracted to pay for the cost of producing it and to yield a profit. One might own a deposit containing gold, but if each ton of rock contains only $1.00 of gold and it costs $2.00 to extract that amount of gold from the ton of rock, then obviously the deposit has no value. But what is not ore today may with improved processes of extraction and transportation be ore tomorrow.

Most of the gold produced today comes from ore that contains from 0.1 to 0.3 ounce, or $2.00 to $6.00 worth of gold per ton of ore, but small deposits containing $10.00 per ton cannot be mined profitably in certain localities.

Silver ores range from 5 to 25 ounces of silver per ton of ore, whereas

most of the copper of the world is obtained from large ore deposits that have less than 40 pounds of copper per ton of ore. Zinc ore must contain from 3 to 30 per cent of zinc; and lead ore from 2 to 10 per cent of lead. Iron ore from the Lake Superior region contains 40 to 60 per cent of iron.

Gold and silver are commonly associated with the other metals, and their presence may enable ore of lower grade than the figures given above to be worked. More than one metal may be won from certain ores; thus lead and zinc, copper and zinc, and silver and lead are common associates.

ORIGIN OF ORE DEPOSITS

When one considers that ore deposits represent concentrations of unusual minerals and that they are sparsely scattered in the Earth's crust, immediately the questions arise: Where did the metals come from, and how did they become concentrated into relatively small and widely scattered bodies; in short, how were the ores formed?

The ore deposits as we see them today occur in diverse forms in all kinds of rocks under conditions which preclude the possibility that they owe their origin to the operation of any one process. We shall first consider their source and next the means by which they have been collected, carried, and deposited in the forms in which we now find them.

Source of the Metals. — The metals must have come from within the Earth, where the igneous rocks originate. They are so widely and intimately associated with igneous rocks or other indications of igneous activity that the two must have had a common origin. Although traces of the metals occur in sedimentary rocks, the sediments themselves were originally derived from igneous rocks, and their metal content likewise is most likely of igneous origin. Of course it must be remembered that ores, like rocks, may be eroded or dissolved and carried elsewhere, and their materials may be formed again in other places. Thus some of them may have obscure parentage.

The conclusion that ores and igneous rocks originate together is further substantiated by the occurrence of traces of most of the metals in the igneous rocks and in hot springs or other emanations of igneous origin.

Collection and Transportation. — The above conclusion of the ultimate source of the metals might lead one to infer that all ore deposits must occur in igneous rocks; but this inference is not correct. Some deposits, it is true, evidently are an original part of an igneous mass, but others fill cracks in igneous rocks and therefore must have been formed after the igneous mass solidified. Moreover, many lie in sedimentary rocks far distant from an igneous mass. Consequently we conclude

that in some cases the metals have been collected within a magma chamber and have remained in the intrusive when it solidified, and in other cases they have been gathered up by some mobile agent, expelled from the magma chamber, and carried some distance to places where deposition of the metals occurred. Thus, some ore deposits are component parts of igneous masses and have been formed at their source by solidification from a magma state, whereas others have been formed as a result of deposition from mobile carriers.

To the first of these we give the name *igneous* or *magmatic* deposits. The substances of which they are composed are more or less common to certain kinds of igneous rocks, and presumably were originally diffused as minute particles of metals or metallic compounds throughout the mother magma. They were collected and concentrated during the cooling of the magma by the process of magmatic differentiation (page 438), and they solidified or crystallized more or less simultaneously with, and as a part of, the original igneous body. Such deposits are, therefore, simply unusual kinds of igneous rock that happen to be of value because their ingredients are desired by man; otherwise they would be looked upon as varieties of rock. Their mode of origin, therefore, is the reason for their name. It must not be supposed, however, that all magmas during their crystallization give rise to these ore deposits. If this were the case such deposits would be common features, whereas they are rare.

The character of the deposits formed in the manner described above will be given attention later. Next, we shall consider the processes by which the metals have been collected and transported from the magma.

The conclusion is inescapable that the carriers of the metals must have been in gaseous or liquid form, depending upon the state of consolidation of the magma at the time they were expelled. This conclusion loses some of its strangeness if we stop to reflect upon the information available regarding such substances.

Highly heated gases are well known to be a part of magmas. In fact, copious exhalations of hot gases take place during volcanic eruptions. Moreover, it is known that they carry metals, since tests made upon such volcanic gases as those of the Valley of Ten Thousand Smokes, and of other places, prove this conclusively. There are also many other observations that lend support to the above statements, but they are too detailed to be considered here. We conclude, therefore, that such highly heated gases have been one important factor in collecting and transporting metals to their present resting places.

Liquids are believed to have been an even more important agent in transporting metals. It is generally thought today that by far the greatest number of ore deposits have been formed by means of hot waters.

Their association with igneous rocks is well known. They are abundantly emitted from volcanoes, and seepages of hot water in the form of hot springs continue in volcanic regions long after eruptions have become quiescent. Furthermore, the rocks adjacent to deep-seated igneous intrusives give evidence that they have been traversed by hot waters that came from the cooling igneous bodies. There is no question, then, that hot waters are normal emissions from cooling igneous bodies.

That hot waters can and do dissolve metals has been demonstrated over and over again in the chemical laboratory. But if doubt remains that natural hot waters are competent solvents and carriers of most of the metals, it is dispelled by the array of tests made upon the hot spring waters of Steamboat Springs, Nevada, and on those of other localities, which show the presence of dissolved metals. Metallic minerals are actually being deposited from some such hot springs. Hot waters are also occasionally encountered in deep mines, and they too contain metals in solution. Hot magmatic waters, therefore, are also carriers of metals from their magma source.

Hot waters follow the gaseous emissions and represent a later and cooler phase of igneous activity. Both gases and waters obtain their load of metals from a cooling magma.

There is still another mode of collection and transportation of metals by liquids. Ordinary meteoric waters, that is, those originally derived from rain, play an important part in the formation and alteration of ore deposits. The commonest example is that of cold surface waters, which dissolve metallic compounds from the upper parts of ore deposits and carry them down beneath the ground water table. Also in the weathering of rocks certain metals, such as iron, are dissolved and carried along by surface waters to places where deposition of their metallic content later takes place. Artesian waters are believed to have dissolved and carried great quantities of lead and zinc, which later were deposited to form the extensive ore deposits of the Mississippi Valley. It is thought, too, that the waters of many hot springs are meteoric waters that have moved downward into hotter regions and have risen again to the surface as hot waters. Such waters also collect and transport metals.

Thus it is seen that as steps in the origin of ore deposits, metals have been concentrated in the magma and deposited by solidification within its crystallized mass; they have been collected from their magma source by highly heated gases and hot waters and carried to the outer cooled portion of the intrusive or into distant rocks; they have been searched out of cold surface rocks by meteoric waters and transported elsewhere.

A further step is necessary, however, to bring about the formation of

ore deposits — the metals must be deposited from their carriers; and we shall now consider the means by which this is accomplished.

Deposition. — It is obvious that the manner of release of the metals from their carriers depends somewhat upon the nature of the carrier. Thus different processes operate to bring about deposition from vapors and gases than from liquids.

Metals are given up from vapors in two ways: (1) by a decrease in temperature and pressure due to contact with cooler rocks; this lowers the solvent power of the vapors and necessitates deposition of minerals; (2) by a chemical reaction between the vapors and the rocks with which they come in contact. This process has already been considered under the heading of Contact Metamorphism (page 354), where it was shown that the hot gases given off by a cooling magma produce profound mineral changes in the adjacent country rock. If the gases contain also notable quantities of metals, *contact-metamorphic deposits* are formed as an incidental phase of the contact metamorphism. The nature of these deposits will be described later.

Deposition from hot waters takes place with or without chemical reaction with the wall rocks, and different kinds of ore deposits result from each process. In the case of chemical reaction, the hot solutions diffuse through or insinuate themselves along cracks or other openings in the rocks, and dissolve all or part of the rock, particle by particle. Simultaneously they deposit equivalent volumes of ore and gangue minerals. Ore replaces country rock, and for this reason the resulting deposits are called *replacement deposits*. The process may be roughly illustrated by supposing that the clay bricks of a wall could be replaced, one by one, by silver bricks; the resulting silver wall would have the same position, volume, and structure as the replaced brick wall. In nature, however, the particles, instead of being the size of bricks, are of molecular size, and the substitution takes place by chemical action in a solution. The substitution may start at a number of centers and give rise to *disseminated deposits*. The centers may enlarge until they coalesce and a large volume of rock is thus more or less completely replaced, forming *massive replacement deposits*. If only the walls of fissures are replaced, *replacement veins* are formed. This replacement process may go on until large ore deposits result. The chemical reaction between gases and rocks mentioned above is also replacement.

If reaction with the wall rocks does not take place, the metals stay in solution until other processes bring about their deposition. In their travel through the rocks they penetrate fissures, joints, pore spaces, caves, or other rock openings. In such places opportunity is afforded for deposition to take place; the cavities become filled with ore and the

resulting deposits are called *cavity-filled deposits.* Deposition in these openings may be brought about by several factors. An important one is the lowering of the temperature of the solutions by their passage through the rocks, making the substances carried less soluble, with the result that they are precipitated as minerals. Or, there may be changes in the concentration of the substances in solution, or the solutions may react chemically with other solutions, gases, or minerals, to bring about deposition.

Deposition of metals from meteoric waters may take place by some of the processes mentioned above, or by evaporation such as takes place when sea water is evaporated and salts are deposited. Organic materials, such as plants and certain forms of bacteria, are also thought to be effective agents in depositing iron from surface waters.

It is thus evident that ore deposits do not owe their origin to any one simple process, and various types of deposits are formed as a result of the operations of the different processes outlined above. Before we take up examples of the different types of deposits, it is desirable to consider briefly some other physical factors that influence ore deposition.

Effect of Temperature and Pressure upon Ore Deposition. — Temperature and pressure play an important part in the character and location of ore deposition. This is particularly true of deposits formed from igneous emanations. During the earlier stages of an intrusion the emanations consist of highly heated gases and vapors, as no liquids can exist at the temperatures that prevail. As they move toward the surface they pass through zones of decreasing temperature and pressure. Gradually they change to hot waters, and deposition of different ores results in response to the changing physical conditions. Certain minerals, such as pyrite, form under a wide range of temperature and pressure, but the deposition of others is restricted to definite ranges of temperature and pressure; consequently they become diagnostic of those particular conditions. Thus, those deposits formed by hot gases and vapors near the contact of the intrusive are characterized by certain associations of minerals of high-temperature origin, such as make up contact-metamorphic deposits (page 439). As the emanations move farther toward the surface, they form deposits which have been divided by Lindgren into three classes:

1. *High-Temperature Deposits.* These are formed at great depth and under high temperature by filling fissures or replacing the country rock. They are characterized by such high-temperature minerals as garnet, pyroxene, amphibole, and magnetite, and contain gold, tin, iron, and copper.

2. *Intermediate Deposits.* These also fill openings, or replace the country rock. High-temperature minerals are absent, and such minerals as quartz, calcite, pyrite, chalcopyrite, galena, and sphalerite are typical of them. The deposits furnish gold, silver, copper, lead, and zinc.

3. *Low-Temperature Deposits.* These have been formed at lower temperatures and shallow depth, and occur chiefly in fissures in shattered rock. They supply most of the gold, silver, and mercury of the world, and are characterized by such minerals as quartz, chalcedony, carbonates, and gold and silver minerals. High-temperature minerals are lacking. Most of the gold-silver deposits of the Rocky Mountain States belong to this group.

Thus it is evident that the environment controls in large part the character of the ore deposits that are found, and the type of deposit found on the present surface depends largely upon the depth to which erosion has reached.

TYPES OF ORE DEPOSITS

From the foregoing section it is evident that ore deposits are not all alike nor are they simple geologic bodies. They have been formed by different processes, are composed of numerous substances, and occur in many forms. Innumerable kinds of ore deposits are the result. A few of the more important types will now be considered.

Primary Deposits

Igneous Ore Deposits. — Since these deposits originate by solidification from magmas, they usually occur in or near intrusive igneous rocks. Their shapes are irregular and they vary greatly in size. They yield magnetic iron ores, corundum, chromium, platinum, nickel, and copper. The deposits seldom consist of masses of pure ore minerals; varying amounts of rock minerals are mixed with them. Bodies of magnetite of course cannot contain much admixed rock mineral, or they will have no value as iron deposits. On the other hand, platinum need constitute only a fraction of one per cent of the rock to make a valuable deposit.

Some ores are usually associated with certain magmas; for example, nickel-copper deposits occur only with a variety of gabbro, and chromium and platinum with peridotite and allied rocks. An area of peridotite is always worthy of search in the hope that chromium or platinum deposits may be discovered in it.

Examples of igneous deposits occur the world over. The immensely valuable magnetite deposits at Kiiruna, Sweden, where 750 million tons of iron ore exist, are of this type, as are also many of the magnetite deposits of the Adirondack region in New York State. The richest nickel deposits of the world, at Sudbury, Ontario, are igneous deposits.

The great platinum and chromium deposits recently discovered in South Africa are also igneous deposits.

Contact-Metamorphic Deposits. — Certain conditions are necessary for the formation of this type of deposit. The intrusion must be deep-seated; the intrusive is usually feldspathic, such as granite or diorite, and valuable deposits occur only where the intruded rocks are limestones or limy shales. It does not follow that all intrusives exert contact metamorphism or that all contact metamorphism is accompanied by ore deposition; in fact, ores are the exception rather than the rule.

Contact-metamorphic deposits usually consist of several bodies of ore irregularly scattered throughout the contact-metamorphic aureole.

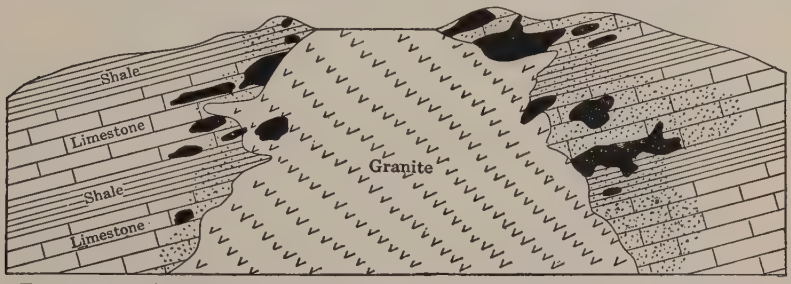

Fig. 305. — Diagram showing a cross section of contact-metamorphic deposits. Stippled area represents contact-metamorphic zone and black, ore.

Most of them, however, lie either adjacent to or quite close to the intrusive (Fig. 305). The ore bodies vary greatly in shape and size. The gangue consists of the metamorphosed rocks described in Chapter XIV, and such minerals as garnet, amphibole, and pyroxene. Magnetite, hematite, and the common sulphides of iron, copper, lead, and zinc are intimately admixed with the silicates. The ore minerals are commonly scattered in small particles throughout the gangue and usually constitute a minor part of the ore. Some of these deposits are worked for iron; others for copper, zinc, lead, gold, tungsten, or molybdenum. They are not as numerous as other types of ore deposits, though individual bodies may be of great value. Usually they are low-grade, and their irregularity of shape, size, and distribution makes them difficult and financially hazardous to mine. They occur in regions where extensive erosion has revealed the larger deep-seated intrusives, and examples are numerous in the eastern and western United States and in Scandinavia. Copper deposits of this type have been mined at Morenci, Arizona; gold deposits at Hedley, British Columbia; lead and zinc deposits at Hanover, New Mexico; iron deposits at Cornwall, Pa., and Banat, Hungary.

Replacement Deposits. — Replacement deposits occur in rocks of all kinds; but limestones are the most common hosts because they are more readily attacked by mineralizing solutions. Replacement deposits are

usually irregular in shape and some of them attain great size (Fig. 306). Many of the world's largest bodies of ore belong to this class. Some deposits consist of disseminated ore minerals sparsely scattered through huge volumes of rock. Such low-grade deposits in the southwestern United States, where they are worked on a large scale, have made this country the world's greatest producer of copper. These are in sharp contrast with the massive deposits in which the replacement process has gone so far that practically no original rock is left within the ore.

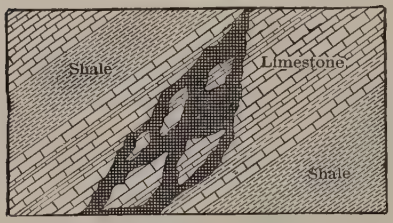

Fig. 306. — Replacement ore deposit. Suspended blocks of unreplaced limestone in the ore (dark) show that the rock was replaced by ore bit by bit, otherwise the inclusion would not be suspended.

The deposits are worked for many metals; vast quantities of copper, lead, and zinc, and considerable amounts of gold and silver are obtained from them.

Examples of replacement deposits are numerous throughout the western and southwestern States. The great silver-lead-zinc deposits of Leadville, Colorado, with a total output of 450 million dollars, are of this type. In the Coeur d'Alene district of Idaho silver-lead and zinc deposits in quartzite have yielded over one-half billion dollars. Each ton of ore, as mined, contains about 8 per cent of lead (160 pounds) and 6 ounces of silver. The large, massive replacement deposits of the Huelva district, in Spain, consist of pyrite with copper, inclosed in slates and porphyry. They have been worked since the time of the Phoenicians and have produced 175 million tons of pyrite with 3 to 4 million tons of copper.

Cavity-Filled Deposits. — There are many kinds of cavities in the rocks, and any of them may become filled with ore, each giving rise to a deposit of different form. However a certain amount of replacement of the walls of cavities has taken place while the cavities themselves were being filled by ore. *Fissures* are the most important class of cavity.

A *fissure vein* in its simplest form is a fissure or fracture in the rock filled with mineral matter deposited from solution (Fig. 307). The fissures may or may not be faults. Fissure veins are numerous, particularly in mountainous regions where rocks are folded or where igneous intrusions have occurred. Their origin by deposition from solutions is commonly shown by coatings or crusts of different minerals parallel to the walls which form sharp boundaries with the country rock (Fig. 308). The central layers may not meet, in which case there remain unfilled

spaces, lined with projecting crystals, called *vugs* (Fig. 308). Many rare and beautiful crystals are formed in these vugs.

There are a great many fissure veins that contain only worthless gangue minerals such as quartz; many also contain insufficient ore minerals to make ore. In fact, those that do contain good ore are relatively few;

Fig. 307. — Fissure vein of gold-quartz; the mining discloses the width of the vein and the wall rock on either side. Cook Mine, Colorado.

otherwise mines would be more numerous than they are. Variation in metallic content constitutes one of the uncertainties of mining operations. There may be good ore in one place along a vein and lean material in another. Those portions in which the ore minerals are more concentrated are called *ore shoots;* a vein usually contains one or more of them. In addition to the actual vein filling, the walls of most fissure veins are impregnated with ore minerals.

Fissure veins may be vertical or inclined, and occur singly, in parallel groups, or in intersecting groups. The intersections intrigue the miner

in the hope of obtaining rich ore. Some fissure veins have been followed several thousand feet along the strike and as far down the dip. The majority, however, are relatively shallow and die out within 2500 feet of the surface. In thickness they range from a few inches to tens of feet, but most of them are less than 10 feet. Fissure veins terminate along their strike and dip by pinching out or abutting against other fissures. Commonly they are cut and displaced by faults, and the finding of the faulted portions is one of the problems in mining operations.

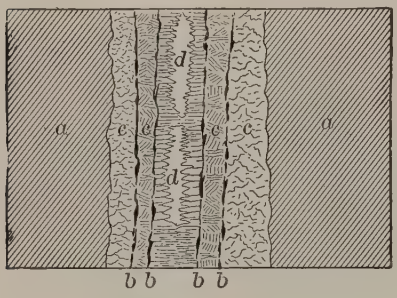

Fig. 308. — Section of a fissure vein; *aa*, wall rock; *bb*, crusts of ore minerals; *cc*, gangue minerals; *d*, *vugs* lined with projecting crystals. Commonly the filling is separated from the wall rock by a thin layer of crushed rock known as *gouge*.

Fissure veins have been enormously productive of silver, gold, copper, mercury, antimony, and other metals. They are world-wide in their occurrence. The veins of Butte, Montana, have yielded from 1882 to the end of 1928 two billion dollars in copper, silver, gold, and zinc; those of California over 600 million dollars in gold; and those of Western Australia over 800 million dollars in gold. A single vein, the Comstock

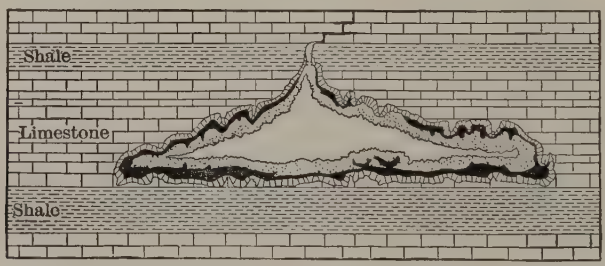

Fig. 309. — Cavity-filled deposit. Shows an open cavity coated by deposited layers of gangue and ore mineral, the latter black. The stalactites and stalagmites of the original opening show that the cavity existed before the deposition. Masses of ore mineral fell from the roof before the last deposit was made.

Lode of Nevada, has contributed nearly 400 million dollars in gold and silver; mercury has been produced from the veins of Almaden, Spain, for over 2000 years.

Other types of cavity-filled deposits have been formed by the filling of caves, joints, bedding planes, cleavage planes, breccia openings, and rock pores. The first three have given rise to the productive lead and zinc deposits (Fig. 309) in Wisconsin. The filling of pore spaces in the vesicular

basalts and interbedded conglomerate of the Lake Superior copper district accounts in large part for the 4 million tons of metallic copper that have been mined from this famous district down to depths of nearly 6000 feet beneath the surface.

Sedimentary Ore Deposits. — In places where sediments are accumulating, certain beds whose component minerals happen to be of commercial value may be deposited from solutions to form ore deposits. Sedimentary iron or manganese ores are formed in this manner (Part II, pp. 164 and 278). Sedimentary iron deposits may be as continuous and as extensive as the sedimentary series in which they occur, and obviously cannot extend beyond its limits. The sedimentary ore beds manifestly partake of the structure of the rest of the sedimentary rocks that inclose them, and may be horizontal, folded, or faulted. Consequently their shape and distribution can be accurately determined by a study of the structure of the region in which they occur.

The Clinton hematite beds, with their 600 million tons of available iron ore, are the most noteworthy representatives of sedimentary ore deposits in North America. They extend from New York State to Alabama and support the great steel industry of Birmingham, Alabama. They have an area of hundreds of square miles, though they are not thick enough to be worked in many places. The ore beds are hard red hematite and average 35 per cent of iron. The iron ores of Lorraine, in France, are another example. These beds are the most important in the world and are estimated to contain 5000 million tons of iron ore.

Disintegration or Secondary Deposits

We have already seen in Chapter III how the rocks become weathered and disintegrated by mechanical and chemical agencies. Ore minerals in the rocks are likewise affected. Some become broken up and are transported as detritus, others are taken into solution and thus carried away, and insoluble ones may remain behind while the materials surrounding them are dissolved and removed. If the valuable materials thus released from the rocks later become concentrated, different kinds of *secondary* or *disintegration deposits* result.

Mechanical Concentrations: Placers. — In this type of deposit nature has operated to produce the results achieved by man when he mines, crushes, and concentrates ore to obtain the desired ore minerals. The ore minerals concentrated in ore deposits or those that are sparsely scattered in the rocks become separated from the surrounding gangue or country rock by mechanical and chemical agencies. The disintegrated materials move slowly down the surface slopes to the nearest stream,

where the water sweeps away the lighter rock and gangue particles, and the heavier ore minerals sink to the bottom or are moved relatively short distances. As thousands of tons of débris are thus moved to the streams, the few ore minerals in each ton of débris will be concentrated in the gravels of the stream bottom until there is accumulated a deposit of sufficient size to be workable. These are called *placer deposits*, and the operation of extracting the valuable minerals is called *placer mining* in contrast to lode mining from bedrock deposits.

Certain conditions are necessary for the formation of placer deposits; the ore minerals must be insoluble, or they will be taken into solution; they must be heavy enough to sink in moving water that will sweep away the same sized particles of rock and gangue; the streams must have sufficient velocity to sweep away the rock and gangue particles. If the stream velocity is great the ore particles also will be swept along until a place is reached where the current slackens; there they will be dropped. If, for example, a quartz vein containing $1.00 of gold in each ton of ore is eroded, the gold particles, being insoluble and heavy, will concentrate in the stream bottom gravels while most of the quartz is swept away. Eventually the gold content of many thousands of tons of ore is thus concentrated. The rich deposits formed in this manner have given rise to the great California gold rush of 1849, the Klondike stampede to the Yukon, and the rich discoveries in Alaska, Australia, and other places. Hundreds of millions of dollars in gold have been extracted from placer deposits.

Relatively few minerals meet the requirements outlined above, for most of them, as will be seen later, are chemically attacked by surface water, or are of low specific gravity. Gold is by far the most common placer mineral, but platinum, tinstone, magnetite, quicksilver, and precious stones also occur in placer deposits.

The placer ore minerals are disseminated in gravels; occasionally pockets or streaks, called *bonanzas*, occur in which a shovelful of gravel may contain a hundred dollars or more in gold. Most placer gold is in the form of fine specks called "dust," but to the joy of the miner larger lumps or *nuggets* are also found, some of which have a value of several thousand dollars each.

The earliest primitive mining undoubtedly was from deposits of this type. The ease of extraction and the richness of some deposits makes them eagerly sought. The hardy miner requires only a shovel and pan to extract the gold; a shovelful of gravel in a pan is dextrously rotated in water until the gravel is washed free from the gold. But this method suffices only for the richer deposits; low-grade gravels are worked by sluicing (Fig. 310); jets of water wash the gold-bearing gravels through

sluice boxes, and the gold collects on riffles or cross bars in the bottom. In ancient alluvial mining, fleeces were placed in the boxes instead of riffles; hence the origin of the fable of the Golden Fleece. More refined methods utilize large mechanically operated dredges that can handle profitably great volumes of gravel containing as little as 8 cents' worth of gold in each cubic yard of gravel. Gold gravels are worked in western North America, Alaska, Yukon, Australia, New Zealand, and Africa. Tin gravels are mined extensively in the East Indies, and most of our

Fig. 310. — Placer mining by sluicing. The gravel containing the specks of gold is washed by powerful jets of water into sluice-boxes, where the gold is caught by riffles. Cariboo District, British Columbia, Canada.

platinum is won from placer gravels, particularly in the Ural Mountains, Russia.

Placer deposits are not restricted to present stream bottoms; the shifting and downcutting of streams has left placer gravels stranded on hillsides and in stream terraces. In arid parts of Australia the wind has served, instead of water, to concentrate the gold, by blowing away the lighter rock matter.

Residual Concentrations. — Relatively insoluble minerals such as manganese, aluminum, or iron, occur scattered through soluble rocks and, in the process of weathering, the rock may be dissolved and carried away, leaving the insoluble particles to accumulate. The process, if long continued, gives rise to residual accumulations of iron, aluminum,

or manganese ore. Similarly, insoluble gold contained in soluble pyrite undergoes residual accumulation.

Chemical Concentrations. — In the process of weathering and erosion some ore minerals are taken into solution, transported short distances, and redeposited elsewhere on or near the surface. Many deposits of bog-iron ore are formed in this manner. Other minerals as well are dissolved and carried down into the original ore deposit and there precipitated, but as this is involved with other changes that affect the upper parts of ore deposits, it will be considered under the following heading.

Superficial Alteration and Enrichment of Ore Deposits

When bedrock ore deposits become exposed at the surface by erosion, they are weathered along with the inclosing rocks. The surface waters and their contained gases oxidize many ore minerals and yield solvents which dissolve other minerals. Thus the upper part of an ore deposit becomes oxidized and leached down to the water table, and the part thus weathered is called the *zone of oxidation.* As the cold, dilute leaching solutions trickle downward through the deposit, they may lose a part of their metallic content in the zone of oxidation, but when they reach the ground water certain of the dissolved metals are precipitated as sulphides to form the *secondary enrichment zone.* The lower unaffected part of the deposit is called the *primary zone.* This zonal arrangement (Fig. 311) is characteristic of weathered ore deposits. In places the secondary enrichment zone may be absent, and rarely the oxidized zone is shallow or lacking, as in glaciated regions. The changes that take place in the different zones will now be considered in more detail and the processes will be illustrated by referring to a simple copper vein consisting of quartz, pyrite, and chalcopyrite.

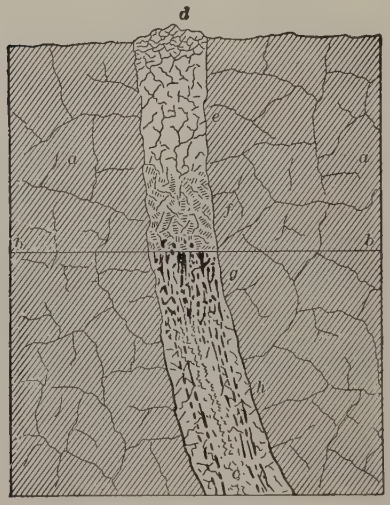

Fig. 311. — Diagram showing zonal arrangement of a weathered vein: *aa*, country rock; *bb*, water table; *cd*, vein; *d–f*, oxidized zone; *d*, capping or gossan; *e*, leached portion; *f*, oxidized ores in oxidized zone; *g*, secondary enrichment zone; *h*, primary zone.

Changes in the Oxidized Zone. — The familiar rusting of iron upon exposure to the weather takes place also in the pyrite of an ore deposit. It becomes chemically altered, and limonite, sulphuric acid and ferric

sulphate, all products of oxidation, are formed. The ferric sulphate is a ready solvent for many ore minerals; it dissolves the copper in the chalcopyrite, forming copper sulphate, which slowly trickles down through the crevices in the deposit. The limonite, however, does not go into solution but remains behind and stains the quartz a rusty color, giving rise to a *gossan* or *iron hat.*

These rusty, cavernous gossans are the surface indications of mineral deposits but usually they do not disclose very much about the mineral character of the ore because the ore minerals have been removed by the leaching solutions. Gossans may result also from worthless pyrite, so that the mere presence of a gossan does not indicate a valuable mineral body. The value usually cannot be determined until at great expense the deposits are penetrated at depth. Certain features of gossans may indicate to the trained geologist something of the original mineral character, but their discussion is beyond the scope of this work. As the oxidation and leaching extend downward usually to the water table, the depth of the leached barren zone may, in arid regions, reach several hundred feet. Because of this, valuable deposits have remained undiscovered for long periods of time.

Silver, zinc, and other minerals also are leached from the oxidized zone. Gold, however, is rarely removed; it remains behind as native gold in the rusty quartz of the oxidized zone. Consequently gold deposits, unlike others, commonly are actually richer in the oxidized zone because they have undergone *residual enrichment.* For example, if primary ore with $5.00 of insoluble gold per ton is one-half removed by solution, there is then left $5.00 in gold in a half-ton, or $10.00 in a whole ton. The ore is thus doubled in value in the oxidized zone by residual enrichment, and contrary to the usual conception, becomes leaner in depth beneath the water table.

The sulphate solutions of the metals formed in the oxidized zone in their journey down through the deposit may lose a part of their metallic content. The copper sulphate, for example, may undergo evaporation and green copper sulphate minerals will be deposited; it may react with calcium carbonate to form the blue (azurite) and green (malachite) carbonates of copper; or with soluble silica to form the light-blue copper silicate (chrysocolla). Native copper also may be deposited. Similarly zinc and lead minerals, native silver, and other minerals of metals are deposited in the oxidized zone. In places these minerals constitute large and valuable ore deposits, but they must always be expected to disappear beneath the zone of oxidation.

Secondary Enrichment Zone. — When the sulphate solutions reach the water level or a place where no oxygen is available, they undergo a

chemical change that causes deposition in the form of secondary sulphide minerals such as *covellite* (CuS) or *chalcocite* (Cu_2S). Thus, to use one metal as an illustration, copper has been taken out of the upper part of an ore deposit and added to a lower part, thereby enriching the lower part in copper and forming the *secondary enrichment zone.* This zone may extend downward several hundred feet but eventually it gradually merges into the unchanged *primary zone* below. This process of leaching above and deposition below may go on continuously as erosion lowers the water level, until primary material that contained only a half of one per cent of copper has been enriched to workable ore containing as much as 3 or 4 per cent of copper. Most of the great copper production of the United States comes from ores that have been thus enriched. If it had not been for the operation of this process, such large mines as those at Bingham, Utah; Santa Rita, New Mexico; Ely, Nevada; Miami and Ray, Arizona; and others, would not exist.

It is evident, therefore, that some weathered ore deposits are barren in the oxidized zone, rich in the secondary zone, and lean in the primary zone; or, in the case of residual enrichment, are richer above than below. The recognition of these surficial changes in ore deposits is one of the modern achievements in geology. It is of great practical importance because it has stimulated deep exploration beneath barren gossans, with the result that many valuable ore deposits have been discovered.

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APPENDIX A

MINERALS

INTRODUCTION AND DEFINITION

Minerals compose the crust of the Earth and are therefore among the most common objects of daily observation. A mineral may be defined as a naturally occurring substance that has a definite chemical composition and a definite combination of physical properties. This eliminates artificial products of the laboratory which may conform to the latter part of the definition. It also eliminates the natural products of organic agencies, as they do not show the definite chemical and physical characters of a mineral.

Minerals are composed of chemical elements. Some consist of single elements, such as diamond and graphite (different forms of carbon), or gold and silver so far as they occur in the free state in nature; but most minerals are made up of two or more elements united in such a way as to give a product that differs in its properties from any of the elements composing it.

At the present time about ninety different elements are recognized, but less than half of them are common and it has been calculated that more than 99 per cent of the crust of the Earth is composed of the following fourteen:

Oxygen, O	49.77 per cent
Silicon, Si	26.09 "
Aluminum, Al	7.34 "
Iron, Fe	4.11 "
Calcium, Ca	3.19 "
Magnesium, Mg	2.24 "
Sodium, Na	2.33 "
Potassium, K	2.28 "
Hydrogen, H	0.95 "
Titanium, Ti	0.39 "
Carbon, C	0.18 "
Chlorine, Cl	0.21 "
Phosphorus, P	0.10 "
Sulphur, S	0.10 "
All others	0.72 "
Total	100.00

Three elements together with their symbols, not included in the above list, are added here because they are present in certain of the minerals to be studied. Each one composes but a small fraction of one per cent of the Earth's crust.

Copper, Cu
Lead, Pb
Zinc, Zn

The number of minerals formed by the combination of even these few elements is very great but the common ones are relatively few.

CHARACTERS OF MINERALS

Chemical Composition

It was said above that minerals have a definite chemical composition. This composition, as determined by analysis, serves to define and distinguish the species, and indicates their relations to each other. Individual minerals react differently to various chemical reagents, and these reactions are one means of determining the kind of mineral under examination. It is beyond the scope of this discussion to treat that aspect of *mineralogy;* but there are many text books that treat the subject fully.

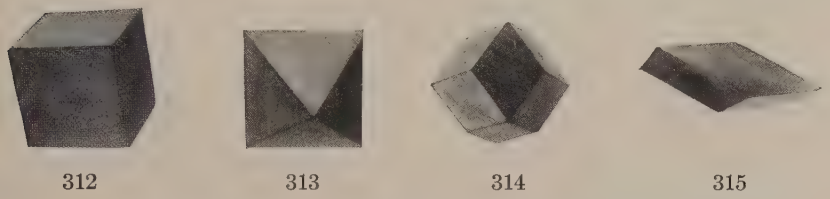
Physical Characters

Structure of Minerals. — Commonly the structure of minerals refers to their outward shape and form. The following descriptive terms are used in this connection, some of which are self-explanatory: *crystallized,* in definite crystals; *columnar; fibrous; botryoidal,* having a group of small rounded forms like a bunch of grapes; *reniform* (kidney-like) and *mammillary,* similar to botryoidal but in larger masses; *foliated* and *micaceous,* occurring in thin sheets; *granular,* in coarse to fine grains; *compact; earthy; stalactitic,* formed in stalactite masses similar to icicles; *massive,* showing compact material with an irregular form; *oölitic,* formed of small, rounded grains which resemble fish roe.

Crystals. — The majority of minerals under favorable conditions will form in crystals. These are bodies which are bounded by plane surfaces that are arranged according to definite laws of symmetry. The division of mineralogy known as crystallography is important and interesting but one whose detailed study takes considerable time. Certain principles will be pointed out here however. Perfect crystals are exceptional things. The great majority of mineral specimens will not show them. When they are to be observed, however, they will help materially in the identification of the mineral. In general the crystals of a certain mineral will

show like or similar habits of crystallization. For instance, the mineral galena, PbS, characteristically crystallizes in cubes (Fig. 312). Magnetite, Fe_3O_4, commonly occurs in eight-sided crystals that are called octahedrons (Fig. 313). Garnets commonly occur in either dodecahedrons (Fig. 314) or trapezohedrons. These crystal habits are characteristic of these minerals and when recognized greatly aid in their identification.

Cleavage and Fracture. — The manner in which some minerals break is characteristic. If the break occurs with a smooth, plane surface the mineral is said to have a *cleavage*. This cleavage always takes place along planes, which may or may not be parallel to crystal faces. Some minerals will show but one cleavage, others two, three, or even six different cleavage planes. The number of planes of cleavage which a mineral shows and their relations to each other help to determine the mineral.



312 313 314 315

Fig. 312. — Model of a cubic crystal.
Fig. 313. — Top of an *octahedron* showing four of the eight faces.
Fig. 314. — Showing six of the twelve faces of a *dodecahedron*.
Fig. 315. — A *rhombohedron*.

Good examples are the cubic cleavage of galena (in three planes at right angles to each other), the rhombohedral cleavage of calcite (three planes not at right angles so that the resulting form is rhombic; Fig. 315), the basal cleavage of mica (in one direction only). If a mineral does not show a cleavage it is said to have a fracture. Various kinds of fracture are as follows: *conchoidal* if the fractured surface is curved like the interior of a clam shell; *fibrous* or *splintery* if it shows a fibrous character; *uneven* or *irregular* if the surfaces are rough.

Color. — The color of a mineral is one of its most conspicuous physical properties. The color of many minerals is a definite and constant property and serves as an important means of identification. For example, the golden-yellow color of chalcopyrite, $CuFeS_2$, the blue-gray of galena, PbS, the black of magnetite, Fe_3O_4, are striking properties of these minerals. However, surface alterations may change the color of a mineral, as is shown in the golden tarnish frequently observed on pyrite, FeS_2. In noting the color of a mineral, therefore, a fresh surface should

be examined. Moreover, many minerals show a variation in color in the different specimens. This may be due to a change in composition such as the gradual substitution of iron for zinc in the mineral sphalerite, ZnS, with the consequent darkening of the color of the mineral; or to impurities such as the red color given to quartz, SiO_2, by the admixture of hematite, Fe_2O_3. Other minerals such as fluorite, CaF_2, show a wide range in color without any apparent change in composition.

Color of Powder or Streak. — *The color of the streak* is an important aid to the identification of minerals. The streak is a thin layer of the powder of the mineral obtained by rubbing it upon an unglazed porcelain plate known as a streak plate. The color of the streak may be similar to the color of the mineral or quite different. For example, some varieties of hematite, Fe_2O_3, have a brilliant black color but give a red-brown streak that positively identifies the mineral.

Luster. — The luster of a mineral is the appearance of its surface due to the manner in which it reflects light. This must not be confused with color for two minerals with the same color may have totally different lusters just as a black paint with a shiny finish, such as an enamel, has an appearance different from that of a black paint with a dull finish because it reflects light differently.

Various descriptive terms are applied to the different kinds of luster exhibited by minerals. A partial list including the more important ones is given below:

Metallic. Having the appearance of a metal. Example, pyrite, FeS_2. (Most of the minerals with a black or dark-colored streak are included here.)

Vitreous. Having the luster of glass. Example, quartz, SiO_2.

Resinous. Having the appearance of resin. Example, sphalerite, ZnS.

Pearly. Having the iridescence of pearl. Example, some varieties of dolomite, $CaMg(CO_3)_2$.

Greasy. Looking as if covered with a thin layer of oil. Some varieties of massive silica, SiO_2.

Silky. Like silk, as the result of a fine fibrous structure. Example, fibrous gypsum, $CaSO_4.2H_2O$.

Adamantine. Having brilliant luster like that of a diamond, C.

Hardness of Minerals. — Minerals vary widely in their hardness, and a determination of this property is often an important aid to their identification. The relative hardness of any mineral may be told by comparing it with a series of minerals that has been chosen as a scale. The scale consists of crystallized varieties of the following minerals, each species being harder than those preceding it in the scale.

Scale of Hardness

1. Talc.	4. Fluorite.	8. Topaz.
2. Gypsum.	5. Apatite.	9. Corundum.
3. Calcite.	6. Orthoclase.	10. Diamond.
	7. Quartz.	

The relative hardness of any mineral in terms of this scale is determined by finding which ones of these minerals it can and which it cannot scratch. In making the determination the following precautions must be observed. Sometimes when a mineral is softer than another, portions of the first will leave a mark on the second which may be mistaken for a scratch. It can be rubbed off, however, whereas a true scratch will be permanent. Some minerals are commonly altered on the surface to material which is much softer than the original mineral. A fresh surface of the specimen to be tested should therefore be used. Sometimes the physical structure of a mineral may prevent a correct determination of its hardness. For instance, if a mineral is pulverulent, granular or splintery in its structure, it may be broken down and apparently scratched by a mineral much softer than itself. It is always advisable when making the hardness test to confirm it by reversing the procedure, that is, by rubbing the unknown on the material of known hardness.

The following materials will serve in addition to the above scale. The finger nail is a little over 2 in hardness, since it can scratch gypsum and not calcite. A copper coin is about 3 in hardness, since it can scratch calcite. The steel of an ordinary pocketknife is just over 5 and ordinary glass has a hardness of 5.5.

Specific Gravity. — The specific gravity of a substance is stated as a number that indicates how many times heavier a given volume of the material is than an equal volume of water. Minerals show a range of specific gravity from about 1.5 to 20. The great majority of minerals range between 2.0 and 4.0. There are various instruments that enable one to determine the specific gravity of a mineral with more or less accuracy, but for ordinary purposes it is sufficient simply to judge the weight of a fair sized piece in the hand. After some practice rather small differences in specific gravity can be detected in this way and a mineral approximately located in respect to this property.

COMMON MINERALS

Only a few of the more common minerals will be described. The student should always compare these descriptions with as many different specimens of the minerals as possible and should note the form, color, and

luster of each sample and make the simple tests that determine the hardness, streak, and specific gravity.

Magnetite

Composition. An oxide of iron, a combination of ferrous and ferric oxides, $FeO.Fe_2O_3$, or Fe_3O_4.

Physical Characters. Color black. Streak black. Hardness 6. Heavy. Strongly magnetic. Usually granular or massive. Occurs in octahedral crystals (Fig. 313).

Occurrence. An important iron ore. It is mined in New York State in the Adirondack Mts., in New Jersey and Pennsylvania, and many other parts of the world. It is common as a minor rock constituent, particularly in the darker colored igneous rocks. The black sand of the sea shore is largely magnetite. It sometimes occurs as a natural magnet, known as a lodestone.

Hematite

Composition. The ferric oxide of iron, Fe_2O_3.

Physical Characters. Color reddish brown to black. Streak light to dark red-brown (Indian-red). Hardness 5.5–6.5. Commonly in botryoidal to reniform shapes with radiating structure. Often earthy. At times micaceous. Rarely in crystals.

Occurrence. Hematite is a widely distributed mineral in rocks and forms the most abundant ore mineral of iron. More than nine-tenths of the iron produced in the United States comes from this mineral. The chief districts lie around the shores of Lake Superior in Michigan, Wisconsin, and Minnesota. Important districts are also located in northern Alabama and eastern Tennessee. Hematite forms the cementing material in red sandstone. It is used also in red paints and as a polishing material, known as rouge.

Limonite

Composition. Hydrous ferric oxide, $Fe_2O_3.H_2O$.

Physical Characters. Color dark-brown to nearly black. Streak yellowish brown. Hardness 5–5.5. Medium heavy. Commonly in mammillary to stalactitic forms with radiating fibrous structure; sometimes earthy.

Occurrence. A minor source of iron. Limonite is a common mineral formed through the alteration or solution of previously existing minerals containing iron. It is found as a cellular mass known as *gossan* in the upper part of sulfide veins; as loose, porous bog-iron ore; associated with siderite as large deposits in limestone and other rocks. It gives the

brown and yellow color to many weathered rocks, sedimentary strata, and soils.

Pyrite

Composition. Iron sulphide, FeS_2.

Physical Characters. Color pale brass-yellow. Streak black. Hardness 6–6.5 (unusually hard for a sulphide). Heavy. Usually granular. Sometimes in crystals, commonly striated cubes or octahedrons.

Occurrence. The most common sulphide mineral. Found in many rocks and is an important vein mineral. Often carries small amounts of gold or copper and so becomes an ore for both these metals. Never serves as an ore of iron but is used as a source of sulphur in the manufacture of sulphuric acid. Its presence in building stones detracts from their value since by its oxidation sulphuric acid is formed, which causes disintegration of the rock and iron oxide which stains its surface.

Chalcopyrite (Copper Pyrites)

Composition. Copper-iron sulphide, $CuFeS_2$.

Physical Characters. Color golden-yellow; often tarnished to bronze or iridescent colors. Streak greenish black. Hardness 3.5 (note difference from pyrite). Heavy. Usually compact massive, rarely in crystals.

Occurrence. A common and important ore mineral of copper. Occurs widely distributed in vein deposits with many other sulphide minerals.

Sphalerite

Composition. Zinc sulphide, ZnS. Nearly always contains a small amount of iron.

Physical Characters. Commonly yellow brown to dark brown in color. Darkens with increase of iron. Resinous to submetallic luster. Hardness 3.5–4. Heavy. White to yellow and brown streak, always a lighter shade than the mineral itself. Perfect dodecahedral cleavage. Usually massive cleavable.

Occurrence. The most common and important source of zinc. Found widely distributed but generally in veins or irregular bodies in limestone. Often associated with galena, pyrite, and chalcopyrite.

Galena

Composition. Lead sulphide, PbS.

Physical Characters. Color lead-gray. Streak grayish black. Hardness 2.5 (soft). Very heavy. Bright metallic luster. Perfect cleavage

in three planes at right angles to each other, forming cubes. Often also in natural cubic crystals (Fig. 312).

Occurrence. The most common and important source of lead. Frequently associated with silver and often serves as an ore of that metal. Also commonly found with zinc minerals.

Calcite

Composition. Calcium carbonate, $CaCO_3$.

Physical Characters. Color usually white or colorless. May be variously tinted, gray, red, green, blue, etc. Usually transparent to translucent. Hardness 3. Light in weight. Perfect cleavage in three planes at oblique angles to each other, giving rhombic-shaped faces (rhombohedral cleavage) (Fig. 315). Often in crystals which generally have a hexagonal cross section. Will effervesce freely on application of a drop of cold acid. This will serve to distinguish calcite from dolomite, $CaMg(CO_3)_2$, another common carbonate, which will not show effervescence under these conditions.

Occurrence. A very common mineral. Is the chief constituent of limestones and marbles. Also a very common vein mineral. Used for the production of lime, plasters, and cement.

Dolomite

Composition. Carbonate of calcium and magnesium, $CaMg(CO_3)_2$.

Physical Characters. Usually white or gray. Sometimes flesh-colored. Transparent to translucent. Hardness 3.5–4 (harder than calcite). Perfect cleavage in three planes not at right angles to each other (rhombohedral cleavage). Light in weight. Vitreous to pearly luster. Will not effervesce upon application of a drop of cold acid unless the specimen is scratched or powdered (differs from calcite). Found in cleavable masses and in crystals which sometimes have curved faces.

Occurrence. Composes rock masses such as dolomite limestone and marble. Also as a vein mineral. Often intimately mixed with calcite. In the rock form, used as a building and ornamental stone, for the manufacture of some cement, and as a source of magnesia for refractory substances.

Gypsum

Composition. Hydrous calcium sulphate, $CaSO_4.2H_2O$.

Physical Characters. Usually white or colorless. Hardness 2 (easily scratched with the finger nail). Light in weight. Has one very

perfect cleavage. May be in tabular diamond-shaped crystals or in cleavable masses. Often also fine granular, sometimes fibrous.

Occurrence. Is a common mineral which is widely distributed in sedimentary rocks, often in thick beds. It frequently occurs interstratified with limestones and shales. Often found in connection with salt beds. Forms twinned crystals similar to Fig. 316. Is chiefly used for the production of plaster of Paris.

Halite (Common Salt)

Composition. Sodium chloride, NaCl.

Physical Characters. Usually white or colorless. Hardness 2.5. Light in weight. Perfect cleavage in three planes at right angles to each other (cubic cleavage). Transparent to translucent. Salty taste. Generally in cubic crystals or in masses showing cubic cleavage.

Occurrence. In extensive beds or irregular masses interstratified with sedimentary rocks and associated with gypsum. Used for culinary and preservative purposes; also very extensively in chemical industry.

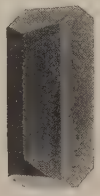
Quartz

Composition. Silicon dioxide, SiO_2.

Physical Characters. Usually colorless or white, but frequently colored by different impurities, yellow, red, pink, amethyst, green, blue, brown, black. Vitreous luster. Transparent to opaque. Hardness 7.

316

317



318

Fig. 316. — Model of a twinned crystal of gypsum.
Fig. 317. — Model of a quartz crystal. This is a six-sided prism terminated by two unequally developed rhombohedrons.
Fig. 318. — Model of an orthoclase crystal.

Light weight. Conchoidal fracture. Commonly in hexagonal crystals similar to Fig. 317. The triangular faces at the ends of the crystals are usually smooth while the rectangular faces about the middle of the crystals are horizontally striated. Also massive.

Varieties. There are many varieties of quartz to which different names are given. A few are as follows: *Rock crystal*, colorless quartz,

commonly in distinct crystals; *Amethyst*, quartz colored purple or violet; *Rose Quartz*, usually massive with a pink color; *Smoky Quartz*, quartz with a smoky yellow to brown or almost black color; *Chalcedony*, finely fibrous material, translucent with a waxy luster; *Agate*, a variegated chalcedony often delicately banded with different colors; *Jasper*, extremely fine grained quartz colored red with hematite.

Occurrence. Quartz is one of the most common minerals. It occurs as an important constituent in many rocks. It is also the most common vein mineral. It makes up the largest part of sands and sandy soils. It is widely used in its various colored forms as ornamental material. It is used for abrading purposes, in the manufacture of glass, porcelain, in paints, scouring soaps, etc. As sand it is used in mortars and cements. Quartzite and sandstone, rocks made up largely of quartz, are used in building, etc.

Garnet

Composition. There are several different garnets which vary from each other in the elements they contain. They are all silicates with somewhat similar formulas. The most common garnet contains ferrous iron and aluminum, $Fe_3Al_2(SiO_4)_3$. Other garnets contain magnesium, calcium, manganese, ferric iron, chromium, etc.

Physical Characters. Color varies with the composition. Most commonly red or brown. May be yellow, white, green, black. Transparent to almost opaque. Hardness 7. Medium heavy. Usually distinctly crystallized, either in a form showing twelve rhombic-shaped faces (dodecahedron, Fig. 314) or twenty-four trapezium-shaped faces (trapezohedron).

Occurrence. Garnet is a common and widely distributed mineral, occurring as an accessory mineral in various kinds of rocks. Used as an inexpensive gem stone and because of its hardness as an abrasive material.

Orthoclase (Potassium Feldspar)

Composition. Potassium aluminum silicate, $KAlSi_3O_8$.

Physical Characters. Colorless, white, gray, flesh-red, more rarely green. Streak white. Hardness 6. Light in weight. Has two good cleavages making 90-degree angles with each other (whence name of mineral). Sometimes in crystals, usually as in Fig. 318.

Occurrence. The most common silicate. Widely distributed as a prominent rock constituent, occurring in many kinds of rocks. Also in large crystals and cleavage masses in what are known as pegmatite veins. From these veins it is quarried in large amounts for use in the manufacture of porcelain.

Plagioclase Feldspars

Composition. Sodium-calcium aluminum silicates.

Physical Characters. Various shades of gray, less commonly white. Transparent to opaque. Hardness 6. Light in weight. Have two cleavages making nearly a 90-degree angle with each other. Commonly distinguished from orthoclase by the color, by the presence on cleavage surfaces of a series of fine parallel striation lines, or by a bluish opalescence. Sometimes crystallize in thin bladed crystals with a curved surface and a pearly luster.

Occurrence. In much the same manner as orthoclase.

Muscovite (Common Light-Colored Mica; Isinglass)

Composition. A complex silicate containing potassium and aluminum.

Physical Characters. Possesses a perfect cleavage in one direction which allows the mineral to be split into excessively thin sheets. The folia are flexible and elastic. Transparent and almost colorless in thin sheets. In thicker blocks, opaque with light shades of brown and green. Hardness 2–2.5. Light in weight. Structure foliated in large to small sheets, sometimes in scales.

Occurrence. A common rock-making mineral. It is found in granite together with quartz and a feldspar and, with the same associations, it occurs in pegmatite veins. Characteristic of a series of rocks made up largely of mica, the minerals being arranged in parallel layers so that the rocks possess a cleavage. These rocks are known as mica schists. Is used chiefly as an insulating material in the manufacture of electrical apparatus. Used as a transparent material in stove doors, etc. There are many other minor uses.

Biotite (Common Dark-Colored Mica)

Composition. A complex silicate containing potassium, magnesium, and aluminum.

Physical Characters. Perfect micaceous cleavage. Folia flexible and elastic. Color usually dark-green and brown to black. Thin sheets usually have a smoky color (differing from the almost colorless muscovite). Hardness 2.5–3. Light in weight.

Occurrence. An important and common rock-making mineral but not as common as muscovite.

Chlorite

Composition. Complex silicate containing magnesium and aluminum. A complex group of minerals of similar characters which are called collectively the Chlorites from their common green color.

Physical Characters. Perfect micaceous cleavage. Folia flexible but not elastic (differing from muscovite and biotite). Color green of various shades. Hardness 2–2.5. Light in weight.

Occurrence. A common rock-making mineral, usually of secondary origin. It results from the alteration of silicates containing aluminum, ferrous iron, and magnesium. To be found where rocks containing such minerals have undergone considerable change due to the heat and pressure to which they have been subjected and have become *metamorphic* rocks. The green color of many rocks is due to the presence of this mineral. This is particularly true of many schists and slates.

Serpentine

Composition. A magnesium silicate, $H_4Mg_3Si_2O_9$.

Physical Characters. Olive-green or yellow-green to blackish-green. Luster greasy or wax-like, silky when fibrous. Hardness 2.5 to 5, usually 4. Light weight. Usually massive but also fibrous or felted.

Occurrence. A common mineral and widely distributed. Always as an alteration product of some magnesian silicate. The massive variety is sometimes used as an ornamental stone. The fibrous variety known as chrysotile is the chief source of asbestos.

Pyroxene and Amphibole

These two common and important rock-making minerals are similar in some respects, and consequently are difficult to distinguish from one another in ordinary rock masses, where good crystal forms do not occur. However it is well to study them separately under favorable conditions, in order to appreciate their differences as well as their points of similarity.

Pyroxene

Composition. A silicate containing chiefly calcium and magnesium; also varying amounts of aluminum, iron, sodium, etc.

Physical Characters. Color usually from light to dark green varying with amount of iron. Also at times nearly white or black. Transparent to opaque. Hardness 5–6. Light in weight. Often in prismatic crystals with eight sides (Figs. 319 and 320). The angle between alternate faces is nearly 90 degrees. These faces will fit into the corner of a box or

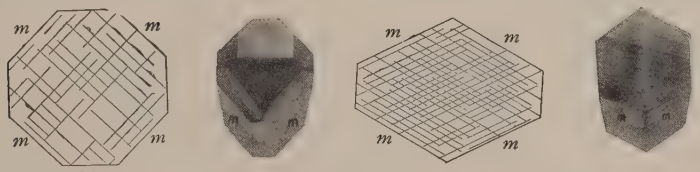
tray. It is by these angles that the mineral can best be told from amphibole. Some specimens show a fair cleavage parallel to the faces lettered *m* in the figures, the angle between the cleavage faces being also nearly 90 degrees.

Occurrence. Pyroxene is a common and important rock-making mineral, being found chiefly in the dark-colored igneous rocks. Seldom found in rocks that contain much quartz.

Amphibole

Composition. Silicate of calcium and magnesium with varying amounts of aluminum, iron, sodium, etc. Similar to pyroxene.

Physical Characters. Color usually light to dark green varying with amount of iron. Also nearly white or black. Transparent to opaque. Hardness 5–6. Light in weight. Often in prismatic crystals with six sides (Figs. 321 and 322). Figure 321 shows that the angles between the



319 320 321 322

Fig. 319. — Cross section of pyroxene normal to the long axis of the crystal. The cleavage traces are parallel to the prism faces (marked *m*). The alternate faces (those marked *m* or those unmarked) make angles of approximately 90° with each other. The cross section has eight sides.

Fig. 320. — Model of pyroxene crystal. Shows the prism faces (*m*) in perspective and a cross section at right angles to the prism zone similar to that in Fig. 319. The inclination of the model somewhat distorts the interfacial angles.

Fig. 321. Cross section of amphibole normal to the long axis of the crystal. The cleavage traces are parallel to the prism faces (*m*) which make angles of 124° and 56° with each other. The cross section has six sides.

Fig. 322. — Model of an amphibole crystal. Shows the prism faces (*m*) in perspective and a cross section at right angles to the prism zone similar to that in Fig. 321. The inclination of the model somewhat distorts the interfacial angles.

faces lettered *m* are 124° and 56° (very different from the corresponding angles in pyroxene). Has a good cleavage parallel to the faces lettered *m*. The differences between the crystals and the cleavage angles in pyroxene and amphibole, and the fact that amphibole has the better cleavage constitute the chief distinctions between the two. Amphibole usually has a higher luster and a smoother surface than pyroxene. In some varieties of amphibole the crystals are long and needlelike, resulting in a fibrous structure. Pyroxene does not occur in this form.

Occurrence. Amphibole is an important rock-making mineral occurring in both igneous and metamorphic rocks, being particularly

characteristic of the latter. Often recognized in the rocks by its elongated bladed structure and good cleavage.

Hornblende is a common dark variety of amphibole.

Pyroxene and amphibole together with biotite are the common dark constituents of nearly all crystalline rocks. The first two can be separated from biotite by the fact that they occur in prismatic crystals that cannot be divided into thin folia; that is, they lack the perfect basal cleavage of the micas. When present as small grains in a rock they lack the high luster characteristic of small brilliant flakes of biotite. They can be told from chlorite by their much greater hardness as well as their form and lack of foliation.

Olivine

Composition. Silicate of magnesium and iron, $(Mg,Fe)_2SiO_4$.

Physical Characters. Usually shades of green, more rarely brownish. Transparent to opaque. Hardness 6.5–7. Vitreous luster. Usually in granular, friable masses, more rarely in perfect crystals.

Occurrence. In many igneous rocks, especially in dark-colored lavas.

APPENDIX B

CHRONOLOGY OF EARTH HISTORY

	Periods	Revolutions (Times of Important Crustal Deformation)	Per cent of Total Time	Dominant Life
CENOZOIC	Quaternary: Recent, Pleistocene		4%	Age of man.
		Cascadian Revolution		
	Tertiary: Pliocene, Miocene, Oligocene, Eocene			Age of warm blooded animals and flowering plants.
		Laramide Revolution		
MESOZOIC	Cretaceous		11%	Age of reptiles and first flowering plants.
	Jurassic			
	Triassic			Age of reptiles and medieval plants.
		Appalachian Revolution		
PALEOZOIC	Carboniferous: Permian, Pennsylvanian, Mississippian		30%	Age of earliest land animals and of the first forests.
	Devonian			
	Silurian			Age of shelled invertebrates and of first fishes.
	Ordovician			
	Cambrian			
PROTEROZOIC	Keweenawan	Some folding and intrusion of granite	25%	Age of primitive invertebrates without shells.
	Huronian			
ARCHEOZOIC		Laurentian Revolution and widespread intrusion of granite	30%	Age of most primitive minute and soft tissued life.
	Timiskaming			
		Folding and local intrusion of granite		
	Keewatin			

Total time represented — 1,000,000,000 ± years.

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